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DEVONIAN OF NEW MEXICO

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ABSTRACT

Three Devonian formations have been recognized in New Mexico—the Canutillo, Sly Gap, and the Percha. New studies now make necessary a revision of, and several additions to, the Devonian system of New Mexico.

The sediments below the Sly Gap formation in the Sacramento and San Andres mountains, hitherto called the "Canutillo formation," from supposed correlation with the Canutillo formation in the Franklin Mountains of Texas, are here considered to comprise the Onate formation.

Time equivalency of the Sly Gap to a portion of the Devonian of Iowa is suggested by new paleontological evidence, the Sly Gap being younger than the Independence shale and slightly older than the Hackberry group.

"Contadero" is a new formational name proposed for a series of carbonaceous shales and gray limestones above the Sly Gap formation near the geographic center of the San Andres Mountains.

The Percha shale consists of two readily divisible members, although no disconformity exists between them. The name "Ready Pay member" is proposed for the "lower Percha," which is composed of black, fissile, nonfossiliferous shale and in most sections involves two-thirds of the total thickness of the formation. "Box member" is proposed for the "upper Percha," which is commonly composed of gray to green calcareous shales with intercalated lenses of limestone. The Box member is confined to the Mimbres Mountains, whereas the Ready Pay member is believed to occur throughout southern New Mexico and probably in the Franklin Mountains of Texas. The Contadero may be a facies of the Ready Pay member.

LOCATION OF THE AREA

Exposures of the Devonian in New Mexico are limited to four major mountain ranges in south-central and southwestern New Mexico (Fig. 1). These ranges are, from east to west, the Sacramento, San Andres, Caballos, and Mimbres (Black) mountains. The structure of the Mimbres Mountains is similar to that of the southern Rockies, whereas the other three ranges represent typical basin-range structures.

The Hueco and Franklin mountains of Texas continue the structural trend of the Sacramento and San Andres mountains of New Mexico. Mud Springs Mountain north of Hot Springs, although separated by the Rio Grande River from the Sierra Caballos, is part of the same structure. Cooks Peak, northeast of Deming, is the result of a laccolithic intrusion and is related to the Mimbres Mountains.

PREVIOUS WORK

The earliest work on the Devonian of New Mexico was confined to the southwestern quarter of the state, where there are numerous outcrops of the Percha Shale. This was done in connection with the extensive development of the ore-bearing Paleozoic sediments in the Mimbres Mountains (Black Range).

Although no formational name was given, the first specific reference to shales of Devonian age was made by C. H.

Gordon and L. C. Graton[1] in 1906. Gordon[2] in 1907 assigned the name "Percha shale" to the Devonian sediments in Kingston–Hillsboro–Lake Valley district (Fig. 1). No specific type section was designated, but the name was applied in general to the many exposures of the

FIG. 1.—Index map of central and southwestern New Mexico

shale along the Rio Percha. Most workers prior to 1907 had combined the Devonian shale with the overlying Mississippian rocks.

The fauna of the Percha shale was described by E. M. Kindle[3] in 1909, much of it being a duplication of the fauna of the Ouray limestone of Colorado, described by G. H. Girty.[4]

The first reference to a fauna of Lime Creek age in New Mexico was made by C. R. Keyes,[5] who reported that the Lime Creek fauna occurs in southwestern New Mexico and southeastern Arizona. For the latter state Keyes undoubtedly had in mind the fauna of the Martin limestone, which had already been correlated with the Hackberry of Iowa by H. S. Williams,[6] a correlation

[1] "Lower Paleozoic Formations in New Mexico," *Amer. Jour. Sci.*, Vol. XXI (4th ser., 1906), pp. 394–95.

[2] "Mississippian (Lower Carboniferous) Formations in the Rio Grande Valley, New Mexico," *Amer. Jour. Sci.*, Vol. XXIV (4th ser., 1907), pp. 60, 62.

[3] "Devonian Fauna of the Ouray Limestone," *U.S. Geol. Surv. Bull. 391* (1909).

[4] "Devonian Fossils from Southwestern Colorado: The Fauna of the Ouray Limestone," *U.S. Geol. Surv. 20th Ann. Rept. (1898–1899)*, Part II, "General Geology and Paleontology," pp. 31–81.

[5] "Lime Creek Fauna of Iowa in Southwestern United States and Northern Mexican Region," *Proc. Iowa Acad. Sci.*, Vol. XIII (1906), pp. 197–98.

[6] F. L. Ransome, "Geology and Ore Deposits of the Bisbee Quadrangle, Arizona," *U.S. Geol. Surv. Prof. Paper 21* (1904), pp. 38, 41.

supported in recent years by A. A. Stoyanow.[7]

Lists of the "Percha shale fauna" from the San Andres Mountains in southern central New Mexico were first made by N. H. Darton,[8] in 1917. Darton, assisted by E. Kirk, pointed out that "the beds in the San Andres Mountains appear to carry a smaller number of typical Ouray forms than are present at Lake Valley, and more forms characteristic of the Martin limestone and Nevada limestone."[9] These workers first recognized that the Devonian fauna in the San Andres Mountains and that of the Percha shale in the Mimbres Mountains are different in age.

M. A. Stainbrook[10] in 1935 suggested the possible existence of a formation or formations in New Mexico of a Devonian age other than the Percha. He presented several problems involved in the correlation of the Devonian strata in the Sacramento Mountains with the Devonian sediments in the San Andres and Sierra Caballos of New Mexico and the Franklin Mountains of Texas. In view of the inadequate field data at that time, Stainbrook's statement of the Devonian problems in New Mexico was remarkably complete. His paper also contained a plate illustrating fourteen genera of various species of fossils collected from a 30-foot vertical exposure of Devonian beds. On the basis of this collection he suggested a correlation with the Independence shale of Iowa.

The Sly Gap formation was named by the writer in 1941.[11] This formation included those sediments in the San Andres and Sacramento mountains previously discussed by Stainbrook.

Investigations made by the writer since 1936 led to the discovery of late Middle Devonian beds below the Sly Gap formation in the Sacramento and San Andres mountains.[12] This discovery prompted further investigations of the Canutillo formation, established by L. A. Nelson in 1937[13] and 1940,[14] in the Franklin Mountains of Texas. But in 1941 the writer did not correlate the Canutillo formation of Texas with the beds below the Sly Gap formation in New Mexico.

The following year G. A. Cooper suggested that the Sly Gap formation may include older Devonian beds of New Mexico.[15] This was a logical suggestion, inasmuch as Stainbrook[16] and L. R. Laudon and A. L. Bowsher[17] had reported the occurrence of *Leiorhynchus* sp. in the Devonian of New Mexico and since a similar *Leiorhynchus* sp. is found in the Canutillo of Texas. To expedite

[7] "Correlation of Arizona Paleozoic Formations," *Bull. Geol. Soc. Amer.*, Vol. XLVII (1936), pp. 486–87.

[8] "A Comparison of Paleozoic Sections in Southern New Mexico," *U.S. Geol. Surv. Prof. Paper 108-C* (1917), pp. 45–46.

[9] *Ibid.*, p. 46.

[10] "A Devonian Fauna from the Sacramento Mountains near Alamogordo, New Mexico," *Jour. Paleon.*, Vol. IX (1935), pp. 709–14.

[11] F. V. Stevenson, "The Devonian Sly Gap Formation of New Mexico," *Oil and Gas Jour.*, Vol. XXXIX (1941), No. 47, p. 65.

[12] Stevenson, "Devonian Formations in New Mexico," *Trans. Ill. State Acad. Sci.*, Vol. XXXIV (1941), No. 2, p. 163.

[13] Abstracts of theses for higher degrees, *Univ. Colo. Studies, Univ. Colo. Bull.*, Vol. XXXII (1937), No. 17, p. 89.

[14] "Paleozoic Stratigraphy of the Franklin Mountains, West Texas," *Bull. Amer. Assoc. Pet. Geol.*, Vol. XXIV (1940), p. 164.

[15] G. A. Cooper *et al.*, "Correlation of the Devonian Sedimentary Formations of North America," *Bull. Geol. Soc. Amer.*, Vol. LIII (1942), p. 1749.

[16] P. 711 of ftn. 10 (1935).

[17] "Mississippian Formations of Sacramento Mountains, New Mexico," *Bull. Amer. Assoc. Pet. Geol.*, Vol. XXV (1941), p. 2141.

description, the writer in 1942[18] correlated the Canutillo formation with the sediments below the Sly Gap in New Mexico. M. A. Fritz[19] conformed to this usage in 1944.

In 1935 K. C. Dunham[20] commented on two distinguishable members of the

writer in 1941[21] and 1942[22] proposed "lower Percha shale" and "upper Percha shale" for the two members of the Percha shale.

Laudon and Bowsher[23] in their report on the Mississippian stratigraphy of the Sacramento Mountains in 1941 used the

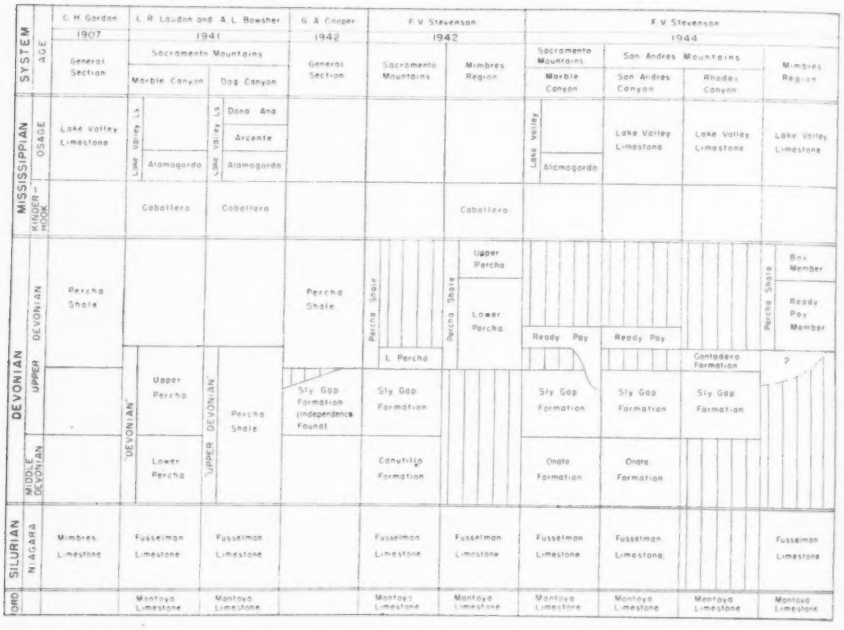

FIG. 2.—Correlation chart of the Devonian of New Mexico

Percha shale, but he did not elaborate on the division of the Percha shale or propose names for the two members. The term "Percha shale" for the Devonian formations in that area, with the subdivision into "upper and lower Percha shale" (Fig. 2). They recognized the complications arising from this usage, as well as the validity of the use of "Percha shale" in this area, as shown by the following quotation: "The recent work of

[18] "Devonian System: The Oil and Gas Resources of New Mexico," *N.M. School of Mines, State Bur. Min. and Min. Res. Bull. No. 18* (2d ed., 1942), pp. 23–24.

[19] "Upper Devonian Bryozoa from New Mexico," *Jour. Paleon.*, Vol. XVIII (1944), pp. 31–41.

[20] "Geology of the Organ Mountains; with an Account of the Geology and Mineral Resources of Dona Ana County, New Mexico," *N.M. School of Mines, State Bur. Min. and Min. Res. Bull. No. 11* (1935), p. 46.

[21] "Devonian Formations of New Mexico" (unpublished Master's thesis, University of Chicago, 1941), pp. 31–32.

[22] Pp. 23–24 of ftn. 18 (1942).

[23] Pp. 2107–60 of ftn. 17 (1941).

Stevenson in the area has demonstrated at least three distinct units in the so-called Percha formation, the upper of which is separated from the others by a distinct angular unconformity."[24]

THE CANUTILLO FORMATION

Varying opinion of stratigraphers regarding the correlation of the Devonian strata in New Mexico and Texas makes necessary the inclusion of a short discussion of the Canutillo formation.

The Devonian strata in the Franklin Mountains of Texas have been known for many years and were considered by Darton[25] as late Devonian ("Percha shale horizon") in age. Darton thought that the Devonian beds in the Franklin Mountains were correlatives of the Devonian sediments to the north, in the Sacramento and San Andres mountains of New Mexico. In a personal communication of Dr. G. H. Girty to Edwin Kirk, quoted in 1940 by L. A. Nelson,[26] the age was designated as "medial Devonian," although no formational name was given to the beds by Girty. Presumably the correlation of these Franklin Mountain beds with those in New Mexico was thus thrown open to suspicion.

L. A. Nelson[27] in 1937 had proposed the name "Canutillo formation" for the Devonian rocks in the Franklin Mountains and indicated their age as medial Devonian. He did not recognize this formation outside of the Franklin Mountains but in 1940 suspected that it may be found in the Hueco Mountains, Texas. He also stated[28] that the Canutillo formation had not been found in New Mexico. Nelson's description of the Canutillo formation is as follows:

> The Middle Devonian is represented in the Franklin Mountains by about 175 feet of sediments consisting of: cherty limestone, light brown color, immediately overlying the Fusselman; a thin bed of fossiliferous gray limestone; a thin bed of dense, almost black sandstone, which weathers brown; and about 40 feet of black fissile shale occurring at the top of the formation.[29]

The writer[30] in 1941 concurred with Nelson that the Canutillo formation does not occur in New Mexico. With the discovery of older Devonian sediments below the Sly Gap formation, however, and the occurrence therein of a species of *Leiorhynchus* similar to a *Leiorhynchus* sp. found in the Canutillo of the Franklin Mountains, and with the consideration of geographic proximity of these two formations, the beds below the Sly Gap in New Mexico were, in 1942 and 1943, correlated with the Canutillo formation.[31] This correlation met with the full approval of L. R. Laudon[32] but was rejected by L. A. Nelson.[33]

Unfortunately, little is known of the meager fauna of the Canutillo formation. It is presumably medial Devonian in age, but the faunal collections made by the writer are not sufficiently diagnostic to make any further statement at present. Study of the Canutillo strata at the type section indicates that they represent near-shore deposits, possibly from brackish water in part, since large plant remains have been found in the upper beds. It is possible that the Canutillo

[24] *Ibid.*, p. 2111.

[25] "Devonian Strata in Western Texas," *Bull. Geol. Soc. Amer.*, Vol. XL (1929), pp. 116–17.

[26] P. 164 of ftn. 14 (1940).

[27] P. 89 of ftn. 13 (1937).

[28] P. 164 of ftn. 14 (1940).

[29] *Ibid.*

[30] P. 163 of ftn. 12 (1941).

[31] Pp. 22–23 of ftn. 18 (1942); "Onondagan Equivalent in New Mexico," *Bull. Amer. Assoc. Pet. Geol.*, Vol. XXVII (1943), pp. 222–23.

[32] Personal communication.

[33] Personal communication.

formation is a time equivalent of the beds below the Sly Gap in New Mexico, even though they are not stratigraphic equivalents. Proof or disproof of this is difficult because of the lack of diagnostic faunas and the slightly different character of the sediments.

THE ONATE FORMATION[34]

"Onate" is a new formational name proposed for the beds below the Sly Gap formation in the Sacramento and San Andres mountains which were formerly identified as the "Canutillo formation." In honor of General Don Juan de Oñate, a late-sixteenth-century explorer and settler in this region, a ridge north of San Andres Peak was named "Onate Mountain." The term "Onate Mountain" was submitted to, and approved by, the Board on Geographic Names in 1944.[35] The type section (Fig. 3)[36] of the Onate formation is located on the north slope of San Andres Canyon, near an abandoned lead mine, in Sec. 18, T. 18 S., R. 4 E., San Andres Mountains, Dona Ana County, New Mexico (Fig. 4).

The Onate formation consists of a variable and intergradational series of shale, siltstone, fine sandstone, and limestone. Identification of these sediments for correlation purposes is complicated by lateral transitional changes, such as gradations from a shale to a siltstone to an arenaceous limestone in a distance of a few hundred feet. Additional difficulties arise from the lack of sufficient index fossils and the absence of strong color contrasts between the beds.

The base of the Onate is easily located, since the older Paleozoic formations, which are massive limestones and siliceous dolomites, are separated by an erosional disconformity from the relatively thin-bedded clastic beds of the superjacent Onate formation.

The top of the Onate is not marked by any disconformity readily detectable in the field. An outstanding difference between the Onate and the overlying Sly Gap formation, however, is the gray-brown color of the former and the tan to light-yellow color of the latter. This striking difference is due, in some degree, to a relatively high percentage of organic material in the Onate and to the occurrence of ferric iron and dolomite in the Sly Gap. Thin, shaly limestone beds, containing many ribbon-like bryozoa, described by M. A. Fritz[37] as *Sulcoretepora anomalotruncata*, occur within 3 or 4 feet of the top of the Onate. These bryozoa are found in most sections and form an excellent horizon marker for the near top of the formation.

The Onate has relatively less shale and more massive beds than the Sly Gap. In addition, flagstone bedding, accompanied by fucoids(?) and worm trails, are prevalent characteristics of the Onate but are rarely found in the Sly Gap. *Leiorhynchus* sp., occurring in great numbers in the Onate, has never been found in the Sly Gap.

The Onate formation is limited in outcrop to the Sacramento Mountains and the southern and central portions of the San Andres Mountains. Strata that may be related to the Onate are found near the northern end of the San Andres Mountains in the vicinity of Mocking-

[34] Pronounced: Own-yah′-tay.

[35] Personal communication. "The following decisions were rendered by this Board on April 13: Onate Mountain; New Mexico, Doña Ana County. A ridge in the San Andres Mountains north of San Andres Peak, between San Andres Creek and Andrecito Creek. T. 18 S., R. 4 E., New Mexico principal meridian. 32°45′ N., 106°34′ W."

[36] The writer introduces a symbol for siltstone in the stratigraphic charts, using a portion of the standard limestone symbol and sandstone.

[37] Pp. 32 and 40 of ftn. 19 (1944).

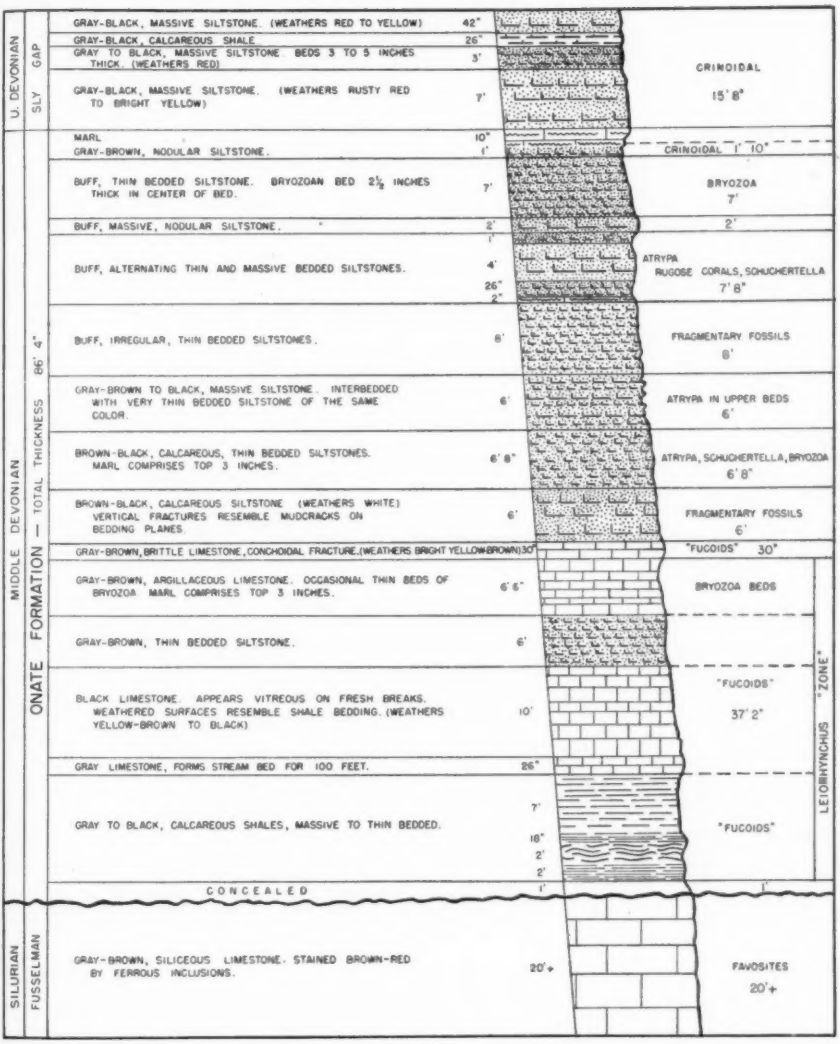
U. DEVONIAN
SLY GAP
GRAY-BLACK, MASSIVE SILTSTONE. (WEATHERS RED TO YELLOW) 42"
GRAY-BLACK, CALCAREOUS SHALE 26"
GRAY TO BLACK, MASSIVE SILTSTONE BEDS 3 TO 5 INCHES THICK. (WEATHERS RED) 3'
GRAY-BLACK, MASSIVE SILTSTONE. (WEATHERS RUSTY RED TO BRIGHT YELLOW) 7'
CRINOIDAL 15'8"
MARL 10"
GRAY-BROWN, NODULAR SILTSTONE. 1'
CRINOIDAL 1' 10"
BUFF, THIN BEDDED SILTSTONE. BRYOZOAN BED 2½ INCHES THICK IN CENTER OF BED. 7'
BRYOZOA 7'
BUFF, MASSIVE, NODULAR SILTSTONE. 2'
2'
BUFF, ALTERNATING THIN AND MASSIVE BEDDED SILTSTONES. 1' 4' 26" 2"
ATRYPA RUGOSE CORALS, SCHUCHERTELLA 7'8"
BUFF, IRREGULAR, THIN BEDDED SILTSTONES. 8'
FRAGMENTARY FOSSILS 8'
GRAY-BROWN TO BLACK, MASSIVE SILTSTONE. INTERBEDDED WITH VERY THIN BEDDED SILTSTONE OF THE SAME COLOR. 6'
ATRYPA IN UPPER BEDS 6'
BROWN-BLACK, CALCAREOUS, THIN BEDDED SILTSTONES. MARL COMPRISES TOP 3 INCHES. 6'8"
ATRYPA, SCHUCHERTELLA, BRYOZOA 6'8"
BROWN-BLACK, CALCAREOUS SILTSTONE (WEATHERS WHITE) VERTICAL FRACTURES RESEMBLE MUDCRACKS ON BEDDING PLANES. 6'
FRAGMENTARY FOSSILS 6'
MIDDLE DEVONIAN
ONATE FORMATION — TOTAL THICKNESS 86' 4"
GRAY-BROWN, BRITTLE LIMESTONE, CONCHOIDAL FRACTURE. (WEATHERS BRIGHT YELLOW-BROWN) 30"
"FUCOIDS" 30"
GRAY-BROWN, ARGILLACEOUS LIMESTONE. OCCASIONAL THIN BEDS OF BRYOZOA MARL COMPRISES TOP 3 INCHES. 6'6"
BRYOZOA BEDS
GRAY-BROWN, THIN BEDDED SILTSTONE. 6'
BLACK LIMESTONE. APPEARS VITREOUS ON FRESH BREAKS. WEATHERED SURFACES RESEMBLE SHALE BEDDING. (WEATHERS YELLOW-BROWN TO BLACK) 10'
"FUCOIDS" 37'2"
"ZONE"
GRAY LIMESTONE, FORMS STREAM BED FOR 100 FEET. 26"
LEIORHYNCHUS
GRAY TO BLACK, CALCAREOUS SHALES, MASSIVE TO THIN BEDDED. 7' 18" 2' 2'
"FUCOIDS"
CONCEALED 1'
1'
SILURIAN
FUSSELMAN
GRAY-BROWN, SILICEOUS LIMESTONE. STAINED BROWN-RED BY FERROUS INCLUSIONS. 20'+
FAVOSITES 20'+

FIG. 3.—Type-section chart of the Onate formation

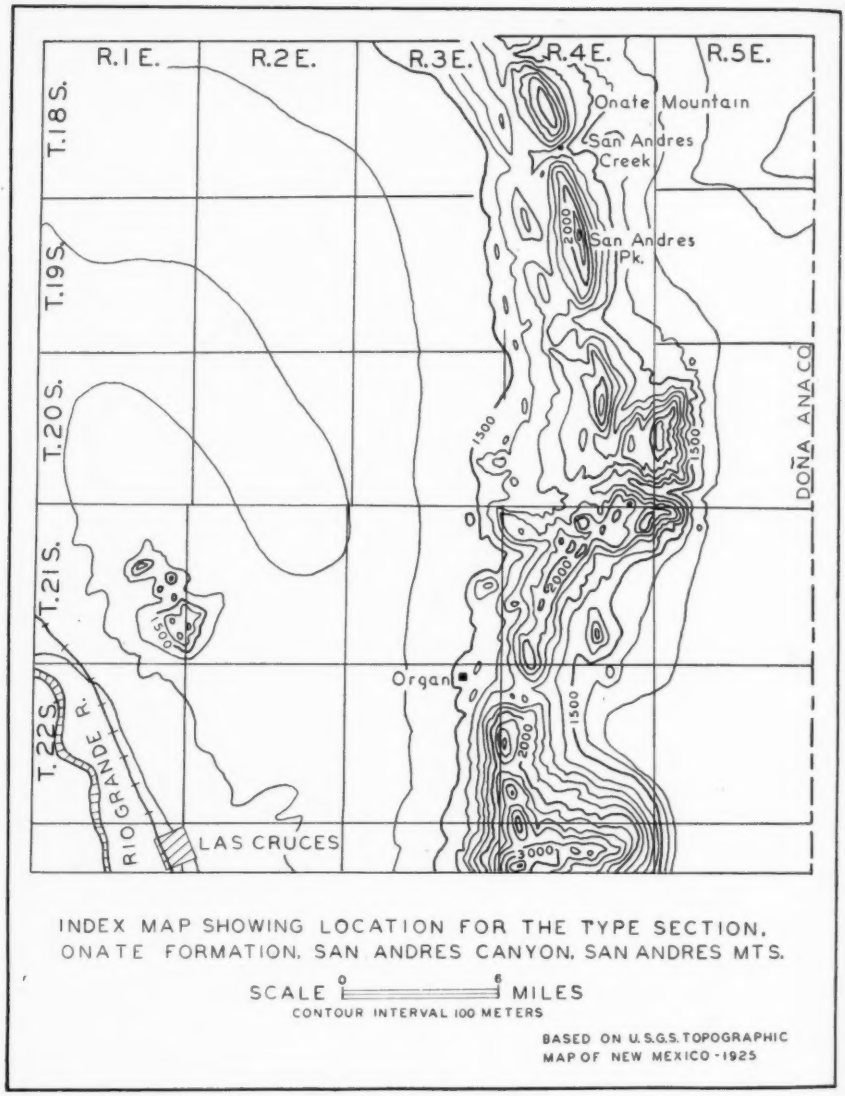
R.1E.
R.2E.
R.3E.
R.4E.
R.5E.
T.18S.
T.19S.
T.20S.
T.21S.
T.22S.
Onate Mountain
San Andres Creek
San Andres Pk.
DOÑA ANA CO.
Organ
RIO GRANDE R.
LAS CRUCES
1500
2000
3000
INDEX MAP SHOWING LOCATION FOR THE TYPE SECTION, ONATE FORMATION, SAN ANDRES CANYON, SAN ANDRES MTS.
SCALE 0 6 MILES
CONTOUR INTERVAL 100 METERS
BASED ON U.S.G.S. TOPOGRAPHIC MAP OF NEW MEXICO - 1925

FIG. 4

bird Gap.[38] Little is known of the character of these Devonian outcrops except that locally there are thin black cherts containing *Tentaculites* and *Paraspirifer acuminatus*, which indicate an Onondagan relationship. A description by Darton[39] of thin cherty limestones (Montoya formation—Ordovician) "overlain by a red shale of unknown age—possibly Percha" in the southern part of the Oscura Mountains, north of Mockingbird Gap, possibly refers, in part, to beds belonging to the Onate formation.

Topographically the Onate formation makes slopes and is covered by a veneer of talus. Where it is exposed on steep grades, the relatively thick beds weather along the bedding planes and have the appearance of man-made steps (Fig. 5).

[38] Stevenson, pp. 222–23 of ftn. 31 (1943).

[39] Darton, " 'Red Beds' and Associated Formations in New Mexico," *Bull U.S. Geol. Surv.*, No. 794 (1928), p. 194.

On the surface of the beds thus exposed, excellent specimens of *Leiorhynchus* sp. can be collected. The underlying formations—Fusselman in the Sacramento Mountains and the Fusselman and Montoya in the San Andres Mountains—are extremely resistant to weathering processes; as a result, the easily eroded Onate and the still less resistant Devonian formations above are commonly expressed in a benchlike indentation on most of the escarpments (Fig. 5).

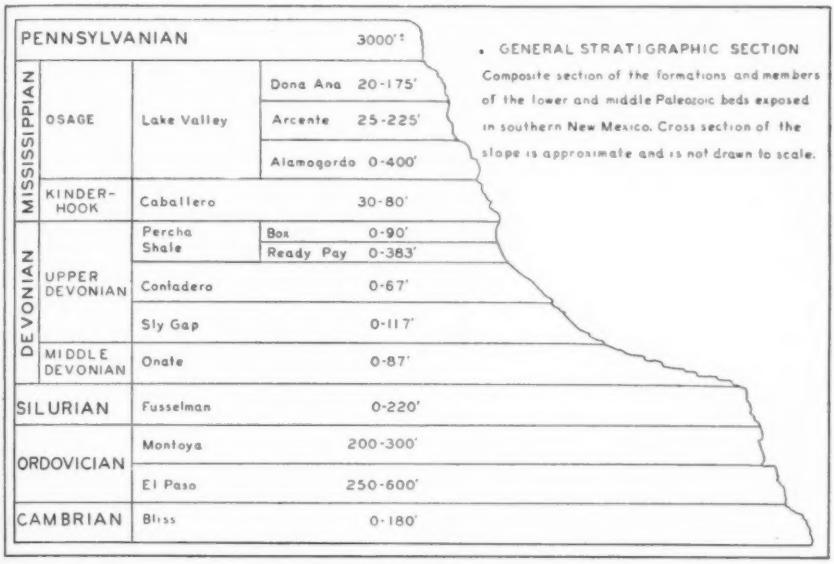

FIG. 5.—General stratigraphic section

The Onate type section (Fig. 3) in the San Andres Mountains exhibits a greater thickness than at any other known locality. The Onate thins to the north, although not so much as the underlying Fusselman limestone. In Rhodes Pass (Canyon) the Fusselman is missing; and the Onate beds, which are absent north of Rhodes Pass, lie unconformably on the Montoya limestone (Fig. 6).

The Onate formation can be traced ap-

proximately 13 miles south of San Andres Canyon, where the Organ Mountains, comprised of extrusives and metamorphosed sediments, disrupt the continuity of the Paleozoic formations. South of the Organ Mountains are bolson plains, succeeded by the Franklin Mountains of Texas, farther south (Fig. 1). due east of Alamogordo; and continuous outcrops of the Onate have been examined from Marble Canyon south to Agua Chiquita Canyon, in the southeast corner of T. 18 S., R. 10 E. In Nigger Ed Canyon, the first canyon south of Agua Chiquita, in the southeast corner of T. 19 S., R. 10 E., the Onate forma-

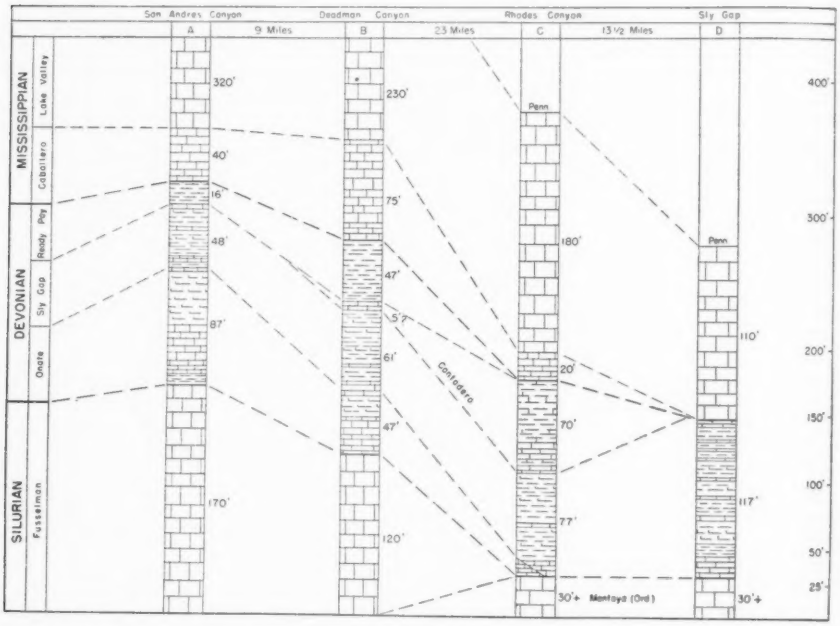

FIG. 6.—South-north correlation chart of the Devonian in the San Andres Mountains

In the Sacramento Mountains the Onate formation averages 35 feet in thickness, although locally there is thickening and thinning of the formation. The Onate is not present in the southern end of the Sacramento Mountains, and it has not been definitely identified toward the head of the Caballero and Gordon tributaries of Alamo Canyon, where the easternmost exposures of older Paleozoic formations are found.

The Onate crops out in Marble Canyon, in the center of T. 16 S., R. 10 E., tion has not been found, although 44 feet of the Sly Gap formation is present.

In Alamo Canyon, which is the first canyon south of Marble Canyon, 4 feet of sandy, basal conglomerate, in which are imbedded small, angular fragments of limestone and quartzite, occur in the Onate formation. This is the only known occurrence of a basal conglomerate in the Devonian of New Mexico.

As pointed out earlier, cross-correlations of the Onate beds are uncertain because of lateral changes. The basal beds

of the Onate in the Sacramento Mountains are usually brownish-gray siltstones or arenaceous limestone, with flagstone bedding. Two exceptions, however, are the basal conglomerate in Alamo Canyon and 3 feet of basal, thin-bedded, calcareous sandstone in Marble Canyon. Another notable variation is the shale which occurs at the base of the Onate in the San Andres Mountains but which has not been found in the Sacramento Mountains.

The fauna of the Onate formation consists of two species of *Atrypa*, a thin-shelled *Strophеodonta*, *Leiorhynchus* sp., *Melocrinites* sp., and a bryozoan described as *S. anomalotruncata*. Other indications of life that are commonly found are fucoids(?), worm trails, and finely disseminated organic material.

Microscopic studies of the Onate sediments have not revealed any ostracodes or other microscopic forms. A thin section of siltstone from the Onate (Fig. 7, *a*) shows a segregation of two sizes of quartz grains, separated by a thin band of organic material. The average size of the quartz grains is 0.03 mm.

No definite age is assigned to the Onate formation, for obvious reasons, although some broad age limits may be drawn from inference and from the available evidence. The beds containing *Paraspirifes acuminatus* from the Mockingbird Gap area may be equivalent in age to the Onate; and, if this is true, the Onate is near-Onondagan in age. In the Devonian of the San Andres Mountains, Darton[40] has reported the occurrence of a variety of *Tropidoleptus carinatus*. Although the exact horizon from which this form was collected is not known to the writer, it is probable that it was collected from what is herein described as the Onate formation. This species generally suggests a Hamilton age; but on the basis of this one form and the distance involved in the correlation, no very definite age can be assumed.

The occurrence of *Melocrinites* sp. is of some significance. This new and undescribed form exhibits specific characteristics found only in advanced crinoids of this age. A similar form, *M. gregeri*, has been reported from the Snyder Creek shale of Missouri; and *M. nodosus* and *M. calvani* from the Cedar Valley and State Quarry limestone of Iowa also bear a strong resemblance to the New Mexican form.

On the basis of evidence at hand, it appears that the Onate is late Middle Devonian or early late Devonian. This opinion is also borne out by the lack of evidence of an erosional period between the Onate and the overlying Sly Gap formation, which is definitely known to be early late Devonian.

Assuming the Onate to be late Middle Devonian, the Devonian beds in the Mockingbird Gap area, which are known to be early Middle Devonian, will perhaps require a new formational name. It is hoped that future studies in the area will permit more specific age determinations and correlation.

THE SLY GAP FORMATION

SAN ANDRES MOUNTAINS

The type section of the Sly Gap formation is located on the south side of Sheep Mountain, in Sly Gap, Sec. 25, T. 11 S., R. 5 E.[41] This area comprises a part of the Greer Ranch, and the section under discussion is located approximately 900 feet in elevation above the north water tank (Fig. 8).

The locality, as defined, was chosen as the most suitable for multiple reasons.

[40] *Ibid.*, p. 194.

[41] Stevenson, p. 65 of ftn. 11 (1941).

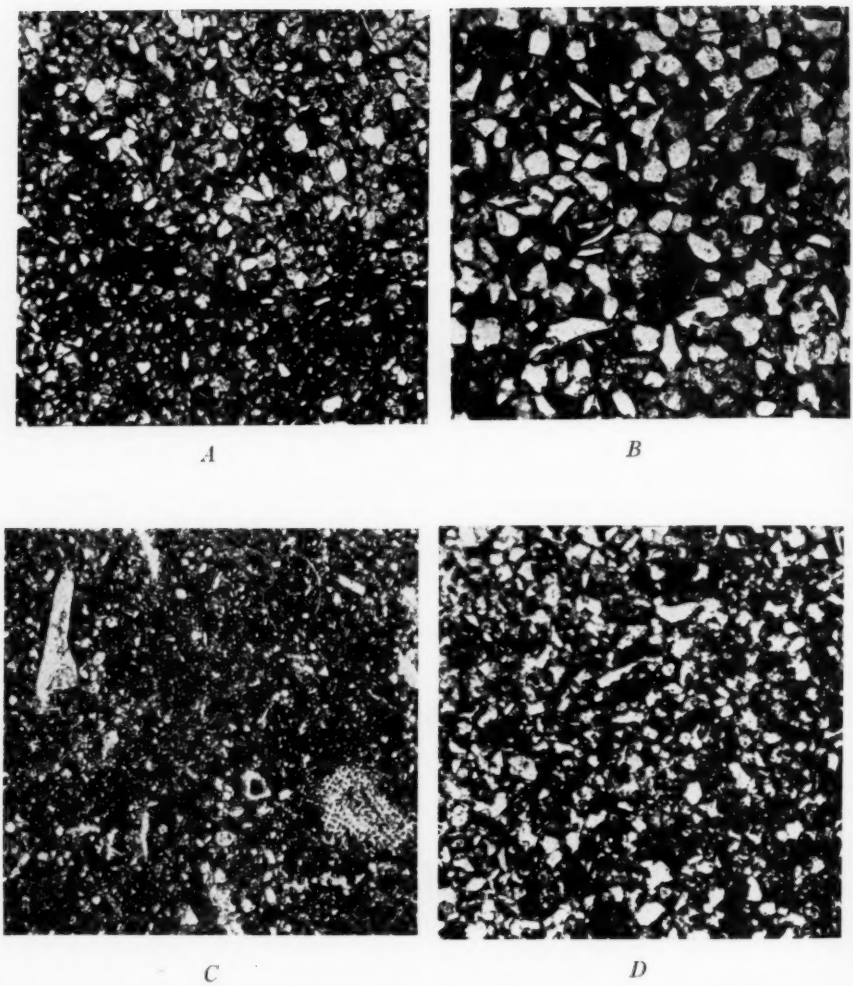

FIG. 7—Microphotographs of Devonian rock: *A*, Onate; *B*, Sly Gap; *C*, Sly Gap; *D*, Sly Gap(?)

In southern New Mexico, ranch roads are annual and must constantly be kept passable. Therefore, localities established within the bounds of a large ranch will usually afford a certain degree of access. The locality also displays excellent weathered slopes for collecting, with a minimum of talus covering. In addition, there are also vertical sections for measurement and study. The fauna is more prolific than at any other locality investigated. Water is available for camping—an important point, inasmuch as it is necessary for anyone wishing to study the type section to expend at least two days in reaching, studying, and leaving the region. It is not advisable to visit this locality if there have been recent rains, as there are several playa areas which are dangerous to cross when wet.

The strata at the type section rest disconformably upon the dark-brown Montoya limestone, of late Ordovician age. Here a green porphyritic dike parallels the contact of the Montoya and the Sly Gap for approximately 100 feet. The Lake Valley limestone rests disconformably on the Sly Gap. The total thickness of the formation at the type section is 114 feet.

The green porphyritic rock, indicated as *B* in Figure 9, is actually a dike, though at the type section it occupies the position of a sill. Many "greenstone" dikes intrude the Paleozoic sections in the southern region of New Mexico, and in many places such dikes parallel various formational contacts for several hundred yards before angling upward through the overlying beds. "Green-

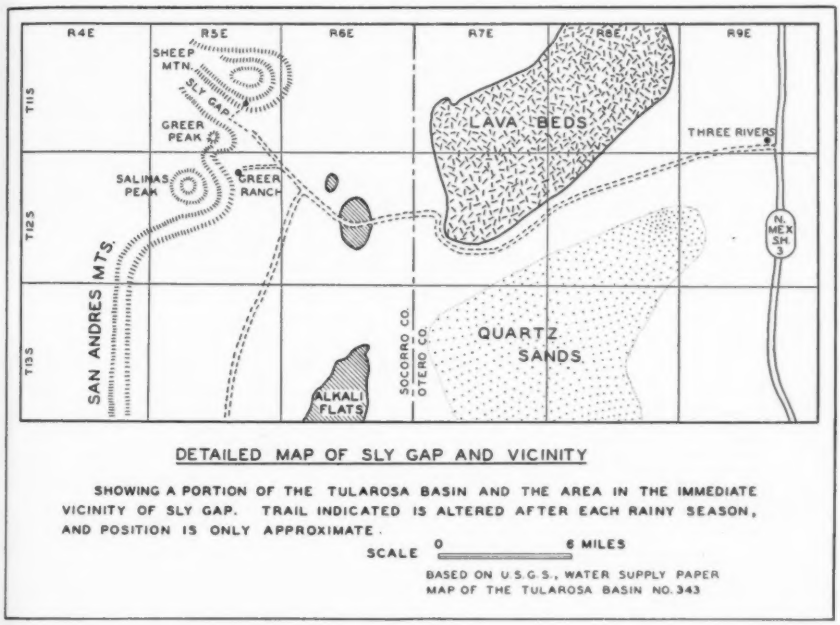

FIG. 8

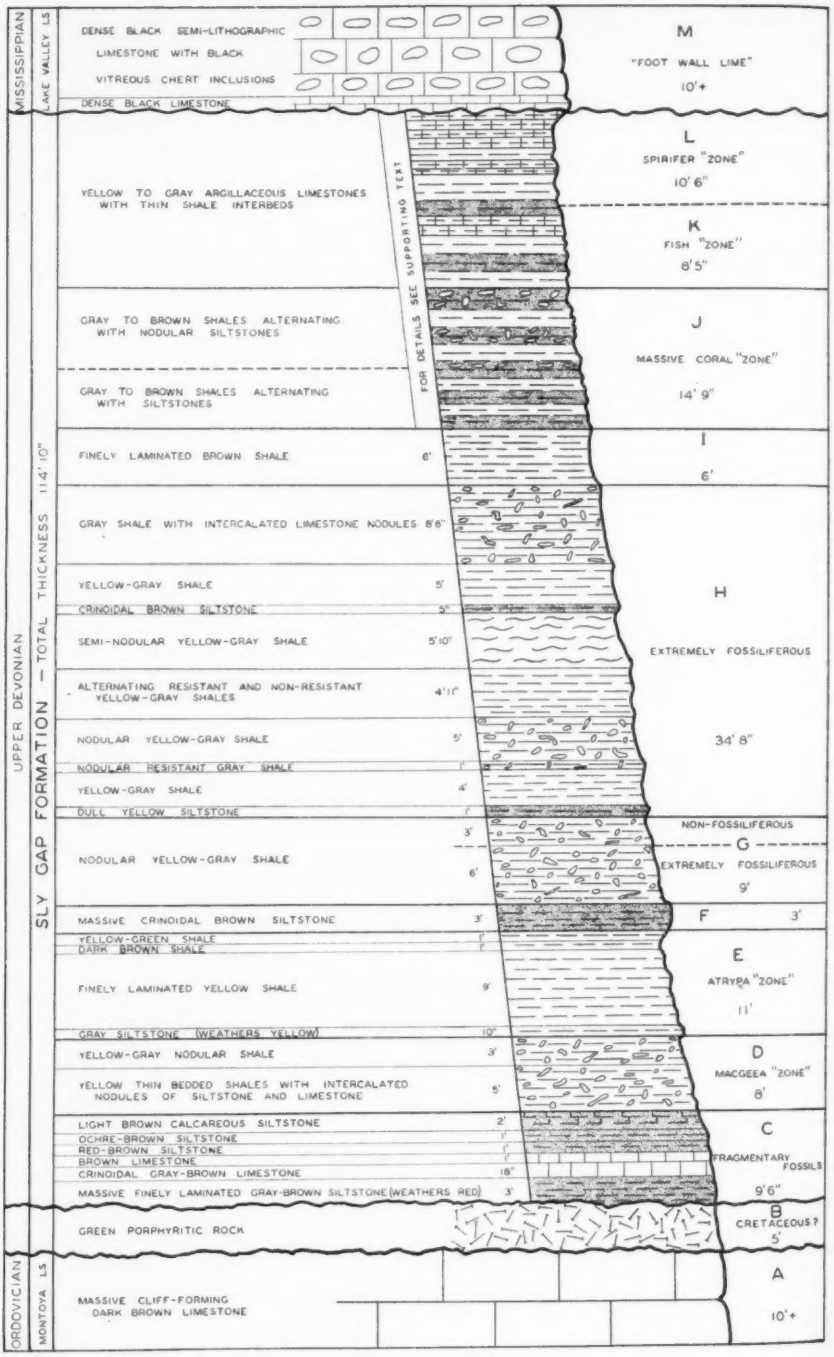

FIG. 9.—Type-section chart of the Sly Gap formation

stone" dikes in the San Andres, as well as in other mountains, have locally been found to mark both upper and lower limits of the Devonian as well as of the Mississippian.

The lithology of the Sly Gap formation makes adequate field description difficult. The formation is composed essentially of thin-bedded, alternating layers of shales, siltstones, and limestones. Rock samples, when tested in the laboratory, show a varying content of $CaMg(CO_3)_2$, $CaCO_3$, SiO_2, and organic material. Beds within the type section are not only very thin but unusually varied in lithology, thus necessitating the measurement of the section with a steel tape, inch by inch. An attempt has been made in Figure 9 to show the changing facies and the apparent effect on the fauna.

Precise "zone" designation by generic names is not entirely satisfactory because the faunules tend to intermingle. Thus, the faunal "zones" established indicate only an abundance of a single genus, which outnumbers all other forms in the same lithologic sequence of beds, and are not to be interpreted as true zones.

Zone *C*, as indicated in Figure 9, is everywhere represented in the Sly Gap formation and makes a red-brown marker for the Paleozoic section of the region where the Sly Gap formation is exposed. Crinoid columnals are commonly found in this horizon, although this characteristic is not so marked in the Sacramento Mountains as in the San Andres.

Wherever possible, there have been indicated on Figure 9, by names and letters, groups of beds carrying a predominance of certain genera. Beds essentially without fossils are indicated as "nonfossiliferous." Beds *G* and *H* are marked "extremely fossiliferous." These beds have many genera represented, without any genus being predominant.

Bed *J* is indicated as the "massive coral zone"; it contains *Alveolites*, *Hexagonaria*, and *Phillipsastrea*. At present this is the only known instance in which any part of the Sly Gap fauna is confined to a small part of the section. *Hexagonaria* and *Phillipsastrea* have been found only in the type section, but *Alveolites* occurs at other localities in the San Andres and Sacramento mountains. Colonial corals have not been found in the Sierra Caballos.

Bed *K*, indicated as the "fish zone," contains a few shark's teeth (see Sly Gap faunal list) and small, black, irregular-shaped nodules of undetermined origin. The strata are not very fossiliferous, and the forms other than those mentioned above are fragmentary.

Bed *L*, indicated as the "Spirifer zone," contains many poorly preserved specimens of *Cyrtospirifer whitneyi*. Fossils other than spirifers collected from this zone have been few. A 3-inch bed of calcareous shale at the top of the "Spirifer zone" marks the contact of the Sly Gap formation and the Lake Valley limestone. "Phosphatic" nodules are found in many places along this disconformable break between the two formations.

A thin section of a rock specimen from the "Spirifer zone" exhibits a field of 60 per cent quartz grains, averaging 0.07 mm. in the longest diameter; 5 per cent primary feldspar; and 35 per cent calcite, with occasional rhombs of dolomite (Fig. 7, *b*). The quartz grains are angular to subangular; well-rounded grains occur rarely. The grain size of the quartz approaches the lower limit of a very fine sandstone, and undoubtedly in places these beds should be called a calcareous sandstone. The hand specimen is sugary in texture; and it is friable, owing to poor cementation. The character of these sediments probably accounts for the poor

preservation of the fauna, as well as for its lack of variety.

For all general purposes, the outcrop area of the Sly Gap formation may be considered as confined to the Sacramento and San Andres mountains and the Sierra Caballo. The Rio Grande River marks the approximate western boundary of this area, and the easternmost exposures in the Sacramento Mountains give the east boundary. The north and south boundaries are within the limits of the aforementioned ranges. The region in which the Sly Gap formation is to be found extends, on the north, from longitude 105°55′ to longitude 107°20′ west; on the west, from latitude 32°30′ to latitude 33°30′ north.

In the San Andres Mountains the Sly Gap formation has been found exposed continuously from a point approximately 13 miles south of San Andres Canyon in the southern end of the San Andres to a point slightly north of Lava Gap in the northern end of the San Andres. The average thickness of the formation is 50 feet in the southern end of the range, and the maximum of 117 feet is found near Sly Gap (Fig. 6).

The Sly Gap formation overlies two different formations in the San Andres Mountains: the Onate, in the southern and central portion, and the Montoya, north of Rhodes Pass (Fig. 6). Two formations are found above the Sly Gap in the San Andres: the Ready Pay member[42] of the Percha shale in the southern and central portion of the range, except where the Contadero formation[43] apparently replaces the Ready Pay; and, at the extreme north end of the San Andres in the vicinity of Sly Gap, the Lake Valley formation, which is found above the Sly Gap formation.

[42] See p. 241.

[43] See p. 239.

Undoubtedly, the best exposures of the Sly Gap are found in the San Andres Mountains. Every canyon and gap (fault-formed) displays clearly the stratigraphic relations and provides fine collecting. The absence of roads and of maps provides the two main handicaps for work in this area.

The Sly Gap formation ordinarily forms an indentation or bench in the lower Paleozoics, similar to the Onate; and in most localities the strata are covered with debris from the overlying Mississippian and Pennsylvanian beds. Where veneering talus is not present, the lighter dun color of the Sly Gap rocks contrasts against the darker beds above and below. Aerial photographs are a great aid in locating new exposures of the Sly Gap because of this contrast in color. They were not used in the Sacramento Mountains but were employed in the northern part of the San Andres and in regions to the west.

SACRAMENTO MOUNTAINS

Exposures of the Sly Gap strata in the Sacramento Mountains display the same general characteristics as those in the San Andres, but it is notable that the sections are thinner in the Sacramentos (averaging 45 feet) and that there is a gradual thinning of the formation to the east and south in this range. Whether post-Devonian erosion or less deposition in Sly Gap time accounts for this thinning is not at present known.

The Sly Gap sediments are well exposed in the Sacramentos; and, with the possible exception of exposures in Rhodes Pass, San Andres Mountains, they are more readily accessible in the Sacramentos. Canyon sections investigated in the Sacramentos are, from north to south, as follows: Marble, Alamo (Fig.

10), Caballero, Gordon, San Andreas,[44] Dog, Escondida, Agua Chiquita, and Nigger Ed canyons.[45]

On the north wall of the first unnamed canyon north of Marble Canyon the Sly Gap formation crops out on the west flank of a northward-plunging anticline continue south in the Hueco Mountains of Texas, but this area has not been investigated by the writer.

Faunal and lithologic zones of the Sly Gap formation are not so consistent throughout the Sacramentos as they are in the San Andres, and with thinning of

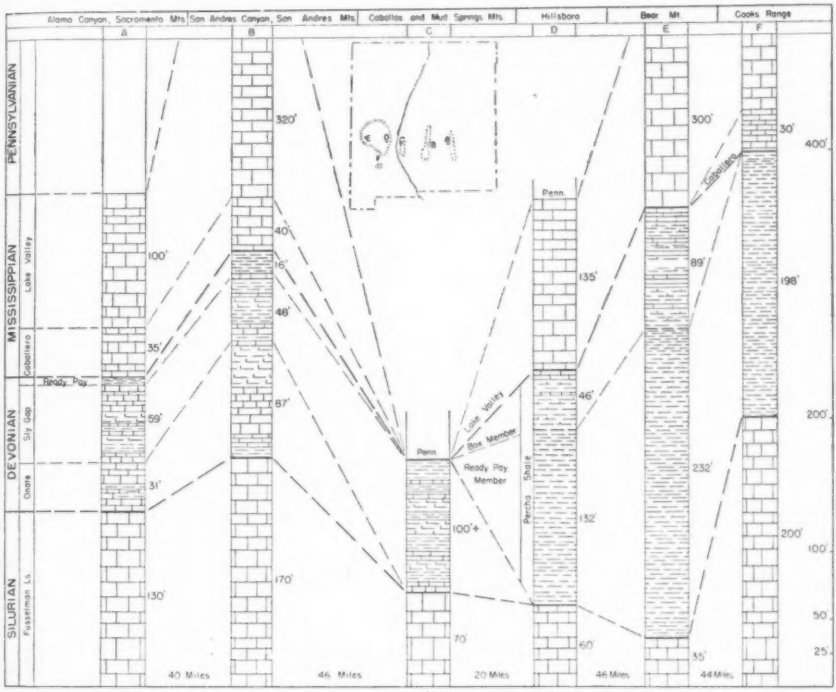

FIG. 10.—East-west correlation of the Devonian in southern New Mexico

whose strike is N. 10° W. This anticlinal structure flattens north of this point, and only Pennsylvanian and younger beds are found beyond it. The Devonian sediments do not crop out south of Nigger Ed Canyon, near the southernmost extension of the mountains. The structural trend of the Sacramentos is known to

[44] Not to be confused with San Andres Canyon, San Andres Mountains.

[45] I. S., Division of Grazing, and Forest Service surveys, 1940.

the formation the stratigraphic units present a somewhat different aspect. More black-shale beds are incorporated in the Sly Gap of the Sacramentos than are found in the San Andres. The probable significance of this feature will be pointed out in the conclusions.

Faunal assemblages collected from various canyons more than a mile east of the western escarpment of the Sacramento Mountains do not display the abundance or the variation exhibited in

the collections taken nearer the escarpment itself. The incomplete collection of Devonian fossils figured in Stainbrook's report[46] on the Devonian of the Sacramentos is from the Sly Gap formation. This collection is valid with the possible exception of a figured specimen of "*Microcyclus?* sp.," which has not been found by the writer in the Sly Gap formation. Similar specimens have, however, been collected from the Caballero formation and identified by L. R. Laudon and A. L. Bowsher[47] as "*Hadrophyllum romingeri.*"

The Devonian section exposed in Dog Canyon presents several puzzling aspects. Locally the Devonian strata are massive and composed of dark-brown, calcareous siltstone approximately 45 feet thick. There is a meager fauna consisting of unidentified fragments and of species of *Schizophoria* and *Leiorhynchus*, which, on the whole, resembles the fauna of the Onate formation more than that of the Sly Gap.

A thin section from the upper beds of the Sly Gap formation in Marble Canyon (Fig. 7, *c*) shows a fine-grained, extremely fossiliferous limestone, with less than 10 per cent quartz grains and rare dolomite rhombs. The strata of the "Spirifer zone" apparently are not represented in the Sacramentos, and the upper strata exposed in Marble Canyon are probably equivalent to the upper beds of Zone *H* of the type section.

A structure of undetermined origin which reveals the only angular unconformity known between the Sly Gap and the overlying Lake Valley formation has been found in Marble Canyon (Fig. 11), which has two main branches—one draining from the northeast and one from the southeast. An elevation, V-shaped in plan with apex pointing almost due west, lies between these two tributaries. The structural feature is shown on the south slope of this elevation in an exposure approximately $\frac{3}{4}$ mile east of the quarry in Marble Canyon. The angular unconformity is found on the west flank of the plunging anticline noted near the beginning of this section. At this place the dip of the Sly Gap beds is 16°, Figure 11 being a view looking northward essentially along the strike.

Out of a total of 48 feet of Sly Gap sediments measured, approximately 45 feet are shown in the area labeled *A* in Figure 11. Not shown in the photograph are 32 feet of the Onate formation, below which the Fusselman limestone lies disconformably. The dip of these older formations is also 16°.

The area marked *B* in Figure 11 shows the Lake Valley strata with a dip of 13° in angular contact with the Sly Gap formation below. The 3° greater dip of the Sly Gap causes a few feet of its topmost beds to pinch out east of the area photographed. The Caballero formation, ordinarily found above the Sly Gap in the Sacramentos, does not crop out in this locality.

In area *C* there are at least 50 feet of black fissile shale, with a few thin stringers of arenaceous shale stained red-brown by oxidation of pyrite. The dip of the black shale varies from a minimum of 28°, near its contact with the overlying Lake Valley, to almost 90°, in its lateral contact with the Onate and Fusselman formations farther down the slope. The strike of the black shale is N. 10° E.

Talus and vegetation obscure the contact with the Fusselman farther down the slope, on the left of the observer. The angular contact of the shale with the Fusselman is probably lost in a relatively short distance, inasmuch as another out-

[46] Stainbrook, p. 712 of ftn. 10 (1935).

[47] Laudon and Bowsher, p. 2124 of ftn. 17 (1941).

crop, perhaps 200 feet to the southwest, exhibits a normal succession of beds without indication of angular unconformity. Except for this very local area,[48] as previously reported, the relation of one Paleozoic formation to another throughout the state of New Mexico is one of parallel disconformity.

Within certain limits the age of all the beds involved in the structure can readily be determined. Since no black shale older than the Sly Gap is known in this area and since the black shale is nonconformable with the overlying Lake Valley, it is clear that the shale is post–Sly Gap and pre–Lake Valley.

The black shale shows no internal slickensides and no slickensided contact with the Sly Gap, Onate, or Fusselman.

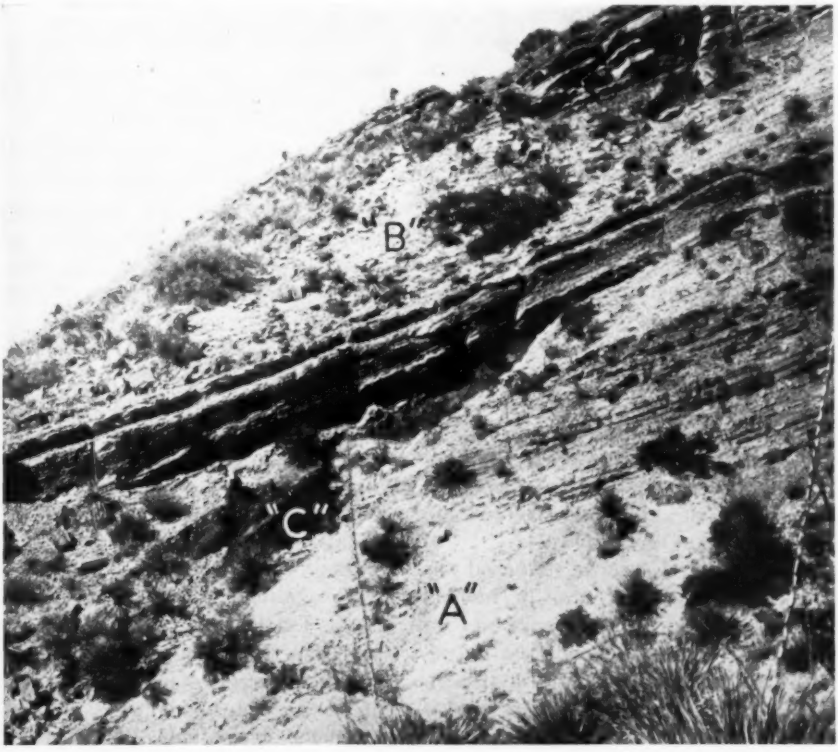

FIG. 11.—Devonian structure in Marble Canyon, Sacramento Mountains

The individual shale beds maintain a uniform thickness regardless of changes in dip or proximity to the adjacent formations. The overlying Lake Valley limestone has not been deformed except for local sagging over the shale. It thus can be definitely stated that the structure is pre-Mississippian. The slight angularity of the Sly Gap and older formations with the overlying Lake Valley

[48] The structure under consideration fails to appear in the cliffs on the north side of the narrow V-shaped (in plan) elevation.

and the eastward tapering-out of the upper beds of the Sly Gap are proof of an erosional period following the tilting of the pre-Mississippian formations. Because the same erosion that beveled the Sly Gap also beveled the black shale, there must have been a relatively thick black shale deposited above the Sly Gap prior to Mississippian deposition. In addition to the 50 feet or more of black shale preserved in this structure, beds of somewhat similar, black, fissile shale, ranging from a few inches to several feet in thickness, have been found above the Sly Gap and below the Mississippian in the Sacramento and San Andres mountains.

Several explanations for the structural relations have been considered, but none is entirely satisfactory. The writer has visited the structure on four occasions, accompanied by L. R. Laudon, J Harlen Bretz, and A. L. Bowsher. The latter participated in its discovery in the summer of 1939. Several others have also visited the area but have been unable to reach a solution to the problem.

One of the first theories advanced was that the black shale records a channel-filling. However, it is impossible to regard the 90° dip of the shale[49] as the result of any known type of initial deposition. Another factor to be considered is the localized tilting of the Sly Gap and older formations. Since this condition is not found at other localities, it would be something of a coincidence if channeling and tilting were to have occurred only once and at the same location.

There is also the possibility that the black shale was deposited against a steep post–Sly Gap erosional surface and that subsequent slumping of the black shale on this surface produced the increased dip. There is no evidence, however, to support this hypothesis, for reasons pointed out above.

It is possible that a cavern had developed in the underlying limestones and that, in a collapse of the roof, the black shale slumped into the cavity. When the very localized area of the structure is considered, this may be the explanation. But it must be pointed out again that the exposed contact of the black shale with subjacent older Paleozoic beds is nearly as straight as if laid out with a tape. Jointing of the limestone is rare, and no appreciable solutional enlargement of such joints as occur in the Sacramentos has been noted. Nor are sinks or caverns known anywhere in the formations under consideration.

Another possible theory invokes submarine landsliding, in which at least all formations from the black shale down through the Fusselman were involved. This is possible but not probable, for such movement should have distorted the shale layers, which would not have retained their uniformity of thickness. Contamination of the shale with fragments of other rock would also be expected. Furthermore, there is little evidence that the Devonian sea bottom was sufficiently irregular to afford the gradient necessary for so much vertical displacement as is indicated in Marble Canyon.

The writer believes that the structure in Marble Canyon was formed through some other process, which permitted the black shale to slump or drop. Normal or high-angle faulting seems most nearly adequate. However, this interpretation demands a fault whose maximum surface trace can hardly exceed 350 feet, while its vertical displacement is at least 140 feet. In the last four years the exposure has been considerably enlarged

[49] The part of the section showing this steep dip did not become exposed until 1942.

by erosion, and it may well be that in the future the talus covering will be removed and further data can be gathered.

MUD SPRINGS MOUNTAIN AND THE SIERRA CABALLOS

The finest exposure in the westernmost extension of the Sly Gap formation is found at Mud Springs Mountain, 2 miles northwest of Hot Springs, New Mexico (Fig. 1), although the section is rather difficult to locate, owing to the folded and faulted complex of the region. The Sly Gap strata here total 48 feet in thickness. They rest disconformably upon the Fusselman limestone and are overlain disconformably by the shale and cherty limestone of the Derry series of the Pennsylvanian system.[50] Much of the Upper Devonian in this area has undoubtedly been removed by pre-Pennsylvanian erosion, as evidenced by abrupt variations in the thickness of the upper shales of the Sly Gap.

The Sly Gap formation is also exposed on the westward-facing escarpment of the Sierra Caballos, south of Hot Springs. The Sly Gap beds found in the Caballos are similar in general character, thickness, and disconformable relations to those exposed in Mud Springs Mountain.

The fauna is the same as that found in the Sly Gap to the east, with the exception of a fossil tentatively identified as *Receptaculites* n. sp., which occurs in large numbers in the Mud Springs locality. The individual fossils are also much larger than those found to the east, but undoubtedly they belong to the same species.

The Mud Springs section is geographically nearer the Percha shale outcrops than any other exposure of the Sly Gap. There is no evidence, however, of any Percha shale or fauna being present. But there is an increase in the amount of shale toward the top of the section, and possibly it can be presumed that the Percha shales were once deposited over this area. The Mississippian limestones that ordinarily have protected the Devonian from excessive erosion were not deposited in this region or were removed by erosion prior to Pennsylvanian deposition.

Due west of the Sierra Caballos, in the vicinity of Hillsboro, New Mexico, are located the easternmost outcrops of the typical Percha shale. Paleozoic formations are not exposed in the Rio Grande River drainage area between the Sierra Caballos on the east and the Hillsboro region to the west; therein lies the difficulty in establishing the relationship of the Sly Gap to the Percha shale. Investigations of the Paleozoic section exposed in the Robledo (Roblero) Mountain, northwest of Las Cruces, did not reveal any Devonian sediments.

PALEONTOLOGY OF THE SLY GAP FORMATION

The only noteworthy studies of the Sly Gap fauna have been made by Stainbrook[51] and Darton.[52] Stainbrook's figured fauna from Gordon Canyon, Sacramento Mountains, includes only fossils found in the Sly Gap assemblage, with the exception of the previously noted "*Microcyclus?*" Darton's list of Devonian fossils from the San Andres Mountains undoubtedly included the fauna of the Onate as well as that of the Sly Gap.

Stainbrook[53] listed, but did not figure, the genera "*Chonetes, Ambocoelia, Lingula, Crania,* and *Liorhynchus* (sic.)" from "Bed 6" in the stratigraphic column,

[50] M. L. Thompson, "Pennsylvanian System in New Mexico," *Bull. N.M. School of Mines*, No. 17 (1942), p. 37.

[51] Pp. 709–14 of ftn. 10 (1935).

[52] Pp. 44–46 of ftn. 8 (1917).

[53] Pp. 710–11 of ftn. 10 (1935).

with which, however, his supporting text fails to conform. Furthermore, the genera listed are not found in the writer's Sly Gap collection or in the collections borrowed. *Ambocoelia* sp. is found in large numbers in the Contadero formation above the Sly Gap in the San Andres Mountains, and it is possible that the specimens of this genus came from Contadero strata. *Leiorhynchus* sp. is found in the Onate formation but has not been found stratigraphically higher. Stainbrook's "Bed 6" occurs beneath the Mississippian and above beds carrying the Sly Gap fauna.

The faunal list on page 239 includes only those genera and species found at Sly Gap itself. No attempt is made to list the complete fauna of this formation, for investigation is still in progress, and a much more detailed descriptive paper will be published as a separate unit. It is significant, however, that the Sly Gap locality does yield approximately 95 per cent of the total fauna, and many forms occurring there are missing at other localities. The faunal richness of the type section may be judged by the fact that, in approximately five hours of collecting, the writer and A. L. Bowsher collected over five thousand specimens from it.

The most diagnostic and consistently present fossils found in the Sly Gap formation are representatives of the genera *Atrypa*, *Productella*, *Nervostrophia* (*Leptostrophia*), *Emanuella?* and *Macgeea*. These five genera are always found in association in every well-exposed section of the formation, and their species occur in larger numbers than do those of other genera.

The genus *Emanuella?* has a cyrtinoid shape and is probably "related to the general group of *Martinia* or *Emanuella*,"[54] with the possibility that the individuals referred to it represent a new genus. This form was figured and identified by Stainbrook[55] as "*Ambothyris?*"

The genus *Macgeea* has been thin-sectioned and is definitely identified as *M. solitaria*, which has been found in the Hackberry of Iowa. From the exterior, *M. solitaria* resembles *M. parva*, which occurs in the Independence shale of Iowa. Stainbrook[56] identified and figured a similar, if not identical, coral as *M. parva*.

Atrypa is the most abundant genus in the Sly Gap fauna and is represented by at least four species: *A. rockfordensis*, *A. devoniana*, and probably two new species.

Both in number and variety the Sly Gap fauna consists predominantly of brachiopods. The corals make up less than 5 per cent of the fossils collected, although *M. solitaria* is one of the five most abundant fossils. One brachiopod specimen from the Sly Gap is identical with *Strophеodonta inflexa*, found in the Snyder Creek formation of Missouri. Other genera, not specifically identified, indicate a close specific affinity to many fossils found in the Hackberry and the Independence shale of Iowa.

Colonial corals from the Sly Gap have been sectioned and are conspecific with *Alveolites rockfordensis* from the Hackberry. Other corals are identified as belonging to the genera *Phillipsastrea* and *Hexagonaria*. Colonial corals have not been reported from the Independence, but they do form an important part of the faunal assemblages of both the Sly Gap and the Hackberry.

The crinoid genus *Arthroacantha* has been found in both the Sly Gap and the Independence shale. The Independence

[54] G. A. Cooper, personal communication.

[55] Pl. 83, Figs. 16–18, of ftn. 10 (1935).

[56] *Ibid.*, Pl. 83, Figs. 8, 25–28.

form was described by A. O. Thomas[57] as *A. mamelonifera;* but the Sly Gap species, though closely related to the Independence form, is new. The occurrence of this rare late Devonian crinoid in the Sly Gap indicates a possible relation to the Independence fauna.

The cephalopod *Manticoceras regulare* has been found in both the Sly Gap and the Hackberry.[58]

The writer believes that the Sly Gap fauna is slightly older than the Hackberry and younger than the Independence fauna. The paleontological evidence also indicates that the fauna of the Sly Gap is slightly older than that of the Martin limestone of Arizona, although many of the genera represented in the two formations are the same. A correlation of the Sly Gap with the Devil's Gate formation of Nevada may also be suggested, at least tentatively.[59]

THE CONTADERO FORMATION

Contadero is a new formational name proposed for a series of carbonaceous shales and limestones above the Sly Gap formation and below the Mississippian strata in the central part of the San Andres Mountains. The type section is located in the south half of Sec. 8, T.-13-S., R.-4-E., near an abandoned mining claim, and in a small tributary draining part of the north slope of Rhodes Canyon. It can be approached on New Mexico Highway No. 52, which passes through Rhodes Pass between the towns of Tularosa and Hot Springs. The type section can be seen approximately 2,000 feet north of the road, and orientation is facilitated by three near-by prospect pits on the slope of the canyon. The name "Contadero" was taken from a Minor Civil Division Map, of the Bureau of Census, issued in 1934. The Contadero political division includes Rhodes Pass and the surrounding area.

The basal gray limestone beds of the Contadero rest without apparent disconformity on the Sly Gap formation, but the gray-black shales and thin, limestone top beds of the Contadero are separated from the overlying Caballero formation by an erosional disconformity (Fig. 12).

Because of the restricted outcrop area of the Contadero, correlations cannot be made to the east or west in the Sacra-

[57] "Echinoderms of the Iowa Devonian," *Iowa Geol. Surv.*, Vol. X (1919–20) (annual reports), pp. 391–506.

[58] A. K. Miller, personal communication.

[59] C. W. Merriam, "Devonian Stratigraphy and Paleontology of the Roberts Mountains Region, Nevada," *Geological Society of America Special Paper No. 25* (1940).

THE SLY GAP FORMATION FAUNAL LIST

BRACHIOPODA
Athyris cf. *vitata*
Atrypa rockfordensis Fenton and Fenton
A. devoniana Webster
Atrypa n. sp.
Aulacella sp.
Camaroloechia cf. *duplicata*
Cranaena cf. *calvani*
Cyrtina hamiltonensis Hall
Cyrtospirifer whitneyi (Hall)
Dalmanella? sp.
Douvillina cf. *arcuata*
Emanuella? sp.
Gypidula cornuta Fenton and Fenton
Hypothyridina cf. *emmonsi*
Hypothyridina n. sp.
Nervostrophia sp.
Productella walcotti Fenton and Fenton
Productella sp. indet.
Pugnoides cf. *calvani*
Schizophoria sp.
Schuchertella cf. *prava*
Spirifer cf. *orestes*
Spirifer cf. *disjunctus*
Stropheodonta inflexa Swallow
Strophonella sp.
Strophonelloides reversus (Hall)
Tenticospirifer? sp.
Tylothyris cf. *mesacostalis*

CRINOIDEA
Arthroacantha n. sp.
Crinoid roots and columnals

ANTHOZOA
Alveolites rockfordensis (Hall and Whitfield)
Aulopora cf. *saxivadum*
Aulopora sp.
Chonophyllum? sp.
Coenites cf. *floydensis*
Hexagonaria sp.
Macgeea solitaria (Hall and Whitfield)
Phillipsastrea cf. *woodmani*
Tabulophyllum? sp.

BRYOZOA
Fenestrellina sp. indet.
Lioclema slygapensis Fritz

CEPHALOPODA
Manticoceras regulare Fenton
Orthoceras sp.

GASTROPODA
Bellerophon sp.
Platyceras sp.
Straparollus sp.

PISCES
Arthrodire cf. *Coccosteus*
Deltodus
Orodus
Petalodus
Ptyctodus

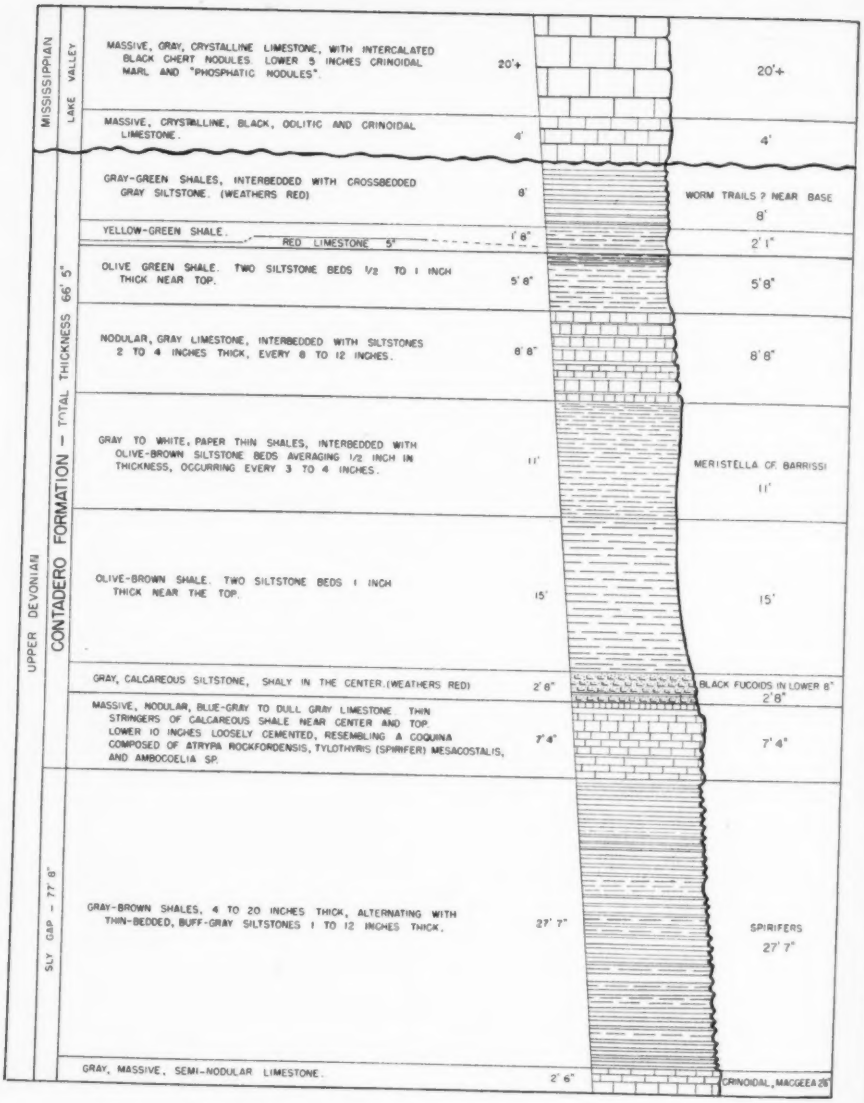
MISSISSIPPIAN
LAKE VALLEY
MASSIVE, GRAY, CRYSTALLINE LIMESTONE, WITH INTERCALATED BLACK CHERT NODULES. LOWER 5 INCHES CRINOIDAL MARL AND "PHOSPHATIC NODULES". 20'+
20'+
MASSIVE, CRYSTALLINE, BLACK, OOLITIC AND CRINOIDAL LIMESTONE. 4'
4'
UPPER DEVONIAN
CONTADERO FORMATION — TOTAL THICKNESS 66' 5"
GRAY-GREEN SHALES, INTERBEDDED WITH CROSSBEDDED GRAY SILTSTONE. (WEATHERS RED) 8'
WORM TRAILS ? NEAR BASE
8'
YELLOW-GREEN SHALE. 1' 8"
RED LIMESTONE 5"
2' 1"
OLIVE GREEN SHALE. TWO SILTSTONE BEDS 1/2 TO 1 INCH THICK NEAR TOP. 5' 8"
5' 8"
NODULAR, GRAY LIMESTONE, INTERBEDDED WITH SILTSTONES 2 TO 4 INCHES THICK, EVERY 8 TO 12 INCHES. 8' 8"
8' 8"
GRAY TO WHITE, PAPER THIN SHALES, INTERBEDDED WITH OLIVE-BROWN SILTSTONE BEDS AVERAGING 1/2 INCH IN THICKNESS, OCCURRING EVERY 3 TO 4 INCHES. 11'
MERISTELLA OF BARRISSI
11'
OLIVE-BROWN SHALE. TWO SILTSTONE BEDS 1 INCH THICK NEAR THE TOP. 15'
15'
GRAY, CALCAREOUS SILTSTONE, SHALY IN THE CENTER. (WEATHERS RED) 2' 8"
BLACK FUCOIDS IN LOWER 8"
2' 8"
MASSIVE, NODULAR, BLUE-GRAY TO DULL GRAY LIMESTONE. THIN STRINGERS OF CALCAREOUS SHALE NEAR CENTER AND TOP. LOWER 10 INCHES LOOSELY CEMENTED, RESEMBLING A COQUINA COMPOSED OF ATRYPA ROCKFORDENSIS, TYLOTHYRIS (SPIRIFER) MESACOSTALIS, AND AMBOCOELIA SP. 7' 4"
7' 4"
SLY GAP — 77' 8"
GRAY-BROWN SHALES, 4 TO 20 INCHES THICK, ALTERNATING WITH THIN-BEDDED, BUFF-GRAY SILTSTONES 1 TO 12 INCHES THICK. 27' 7"
SPIRIFERS
27' 7"
GRAY, MASSIVE, SEMI-NODULAR LIMESTONE. 2' 6"
CRINOIDAL, MACGEEA SP.

FIG. 12.—Type-section chart of the Contadero formation

mentos or Sierra Caballos. In cross section from north to south in the San Andres Mountains the Contadero formation roughly resembles the shape of a shallow bowl (Fig. 6).

Designation of the Contadero in this paper as a formation does not preclude the possibility of other stratigraphic terms eventually being applied to these beds. The Contadero formation may prove to be a tongue, member, or lentil of the Sly Gap formation; or it may be the age equivalent of the Ready Pay member of the Percha shale. Nevertheless, the name "Contadero" will retain its usefulness as designating a mappable and distinct lithologic unit, regardless of any subsequent change in its stratigraphic classification.

The fauna of the Contadero is, in general, the same as that of the Sly Gap, and undoubtedly many of the species are conspecific. An undescribed species of *Ambocoelia* which has not been found in the Sly Gap, however, occurs abundantly in the basal beds of the Contadero. In many basal sections the *Ambocoelia* are so abundant that the rock is essentially a coquina.

In the type of preservation of the fossils, however, the contrast seen in the two faunas is as great as the differences in the lithology of the Sly Gap and the Contadero. The Contadero fauna is preserved in gray argillaceous limestone with layers of white calcite representing the actual shell structure, whereas the Sly Gap fossils have a center of buff, dolomitic limestone and, although similarly layered with calcite, the dolomite matrix gives a yellow-buff color to the specimens.

THE PERCHA SHALE

The Percha shale is best exposed in the region between Lake Valley, Hillsboro, and Kingston and in the vicinity of Santa Rita and Silver City. There are also many outcrops of the Percha between these two regions. It was the only recognized Devonian formation in New Mexico from 1907 through 1940, inclusive (Fig. 2), although the presence of other Devonian formations was suspected.

The specific type section of the Percha shale has not been certainly identified. The type locality was designated by Gordon[60] as on Percha Creek in the vicinity of Hillsboro, New Mexico. In this area very few sections of the Percha shale expose the basal and upper contacts, but an excellent section (Fig. 13), which does show both contacts, is found $2\frac{1}{2}$ miles southeast of Hillsboro in the SW.$\frac{1}{4}$, SW.$\frac{1}{4}$, SE.$\frac{1}{4}$, of Sec. 14, T. 16 S., R. 7 W.[61] This exposure is approximately $\frac{1}{2}$ mile south of the narrow canyon called "The Box," through which the Percha Creek flows eastward into the Rio Grande. This is the section that subsequent workers have probably referred to as the type, and the writer proposes that it be adopted as the neotype section. The area, however, is heavily mineralized, and the fossils occurring here are not so well preserved or so abundant as in other areas.

In 1942 the Percha shale was divided into the "lower Percha shale" and the "upper Percha shale"[62] (Fig. 2). But, because of possible future complications from this nomenclature, the writer proposes the name "Ready Pay member" for the "lower Percha shale" and "Box member" for the "upper Percha shale." The Ready Pay member was named after a gulch of that name which drains to the south into Percha Creek, 400

[60] Pp. 60 and 62 of ftn. 2 (1907).

[61] U.S. Geol. Surv., New Mexico Hillsboro Quadrangle (1940 ed.).

[62] Stevenson, pp. 23–24 of ftn. 18 (1942).

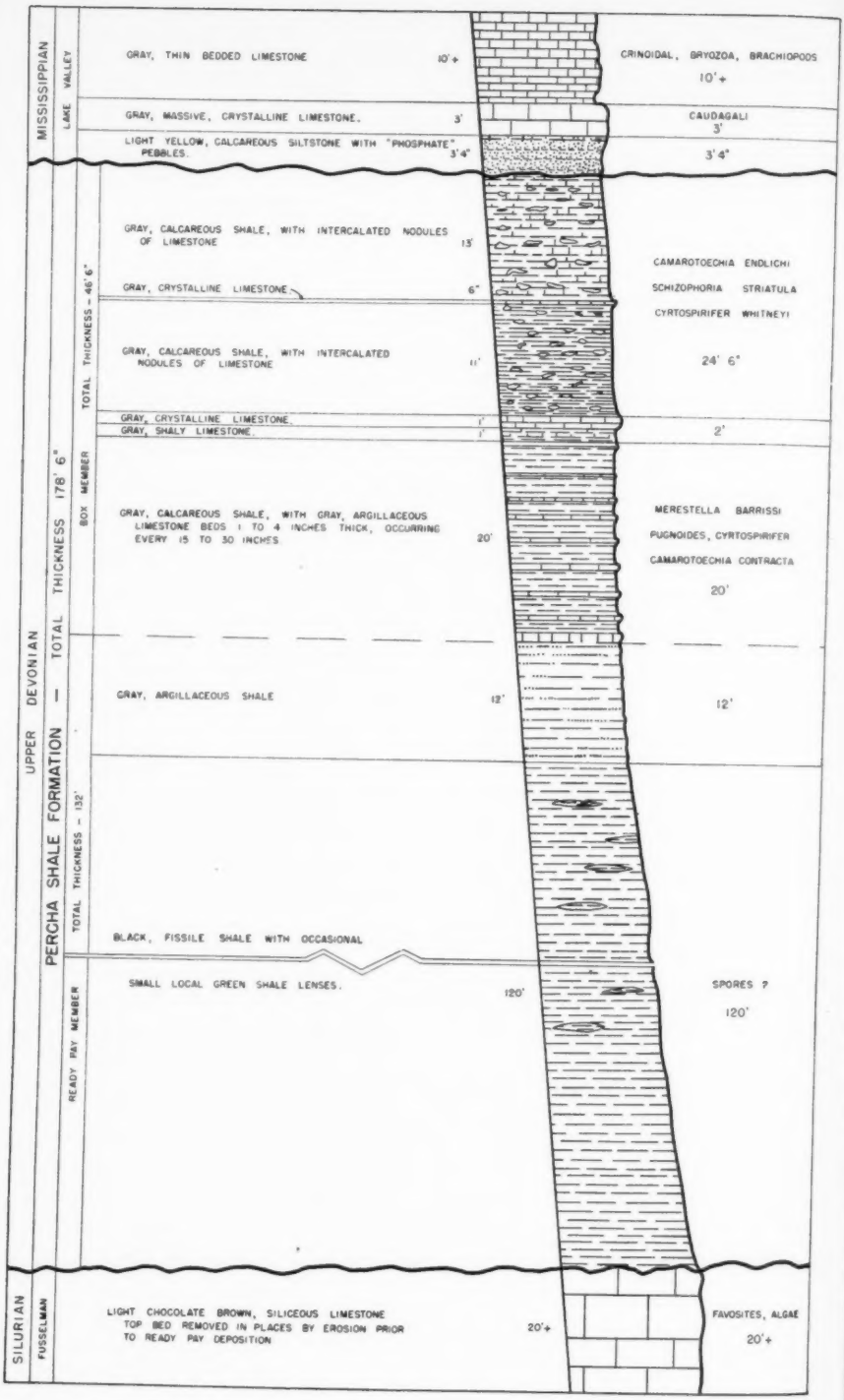
MISSISSIPPIAN
LAKE VALLEY
GRAY, THIN BEDDED LIMESTONE 10'+
CRINOIDAL, BRYOZOA, BRACHIOPODS 10'+
GRAY, MASSIVE, CRYSTALLINE LIMESTONE. 3'
CAUDAGALI 3'
LIGHT YELLOW, CALCAREOUS SILTSTONE WITH "PHOSPHATE" PEBBLES. 3'4"
3'4"
UPPER DEVONIAN
PERCHA SHALE FORMATION — TOTAL THICKNESS 178' 6"
BOX MEMBER TOTAL THICKNESS - 46' 6"
GRAY, CALCAREOUS SHALE, WITH INTERCALATED NODULES OF LIMESTONE 13'
GRAY, CRYSTALLINE LIMESTONE 6"
CAMAROTOECHIA ENDLICHI
SCHIZOPHORIA STRIATULA
CYRTOSPIRIFER WHITNEYI
GRAY, CALCAREOUS SHALE, WITH INTERCALATED NODULES OF LIMESTONE 11'
24' 6"
GRAY, CRYSTALLINE LIMESTONE. 1'
GRAY, SHALY LIMESTONE. 1'
2'
GRAY, CALCAREOUS SHALE, WITH GRAY, ARGILLACEOUS LIMESTONE BEDS 1 TO 4 INCHES THICK, OCCURRING EVERY 15 TO 30 INCHES 20'
MERESTELLA BARRISSI
PUGNOIDES, CYRTOSPIRIFER
CAMAROTOECHIA CONTRACTA
20'
GRAY, ARGILLACEOUS SHALE 12'
12'
TOTAL THICKNESS - 132'
BLACK, FISSILE SHALE WITH OCCASIONAL
READY PAY MEMBER
SMALL LOCAL GREEN SHALE LENSES. 120'
SPORES ?
120'
SILURIAN
FUSSELMAN
LIGHT CHOCOLATE BROWN, SILICEOUS LIMESTONE TOP BED REMOVED IN PLACES BY EROSION PRIOR TO READY PAY DEPOSITION 20'+
FAVOSITES, ALGAE
20'+

FIG. 13.—Type-section chart of the Percha shale formation. (The zigzag double line through the Ready Pay member represents shale not included to scale.)

yards east of "The Box," for which the upper member was named.

The Ready Pay member is composed of black, fissile, nonfossiliferous shale; and in most sections it comprises two-thirds of the total thickness of the Percha shale. The Ready Pay member grades without any marked break into the Box member. The Box member is composed of gray to green calcareous shales with intercalated lenses and nodules of limestone.

The Box member becomes progressively more calcareous to the west of Hillsboro; and on Bear Mountain, northwest of Silver City, interbedded in the shale are many massive beds of limestone, one attaining a maximum of 16 feet. This contrasts strongly with the scattered limestone nodules found in the Box member in the region of Hillsboro and Lake Valley. The Box member, which has not been recognized outside of the Mimbres region, carries the entire Percha fauna.

The Ready Pay member is believed to occur throughout southern New Mexico and probably in the Franklin Mountains of Texas. Black, fissile, nonfossiliferous shale has been found above the Sly Gap and below the Mississippian formations in the Sacramento and San Andres mountains. As previously pointed out, the Contadero formation, whose upper portion consists mainly of shales, has, in a few localities in the San Andres, supplanted the Ready Pay (Fig. 6). In Mud Springs and Caballos mountains the upper part of the Sly Gap becomes increasingly shaly, with black shale predominant at the top. In Cooks Peak northeast of Deming, New Mexico, 182 feet of black fissile shale is found below the Caballero formation. The Box member is recognized a short distance to the north but has not been detected in Cooks Peak. In the Franklin Mountains of Texas the Canutillo formation has approximately 40 feet of black, fissile, nonfossiliferous shale at the top and is overlain by the Helms formation (Mississippian). The shale may or may not be stratigraphically equivalent to the Ready Pay member.

There is no definite proof that the widely occurring black shales of southern New Mexico are correlatives of the Ready Pay, but the stratigraphic position of these shales and the general Paleozoic history of the region indicate their probable contemporaneity. This view is strengthened by the discovery of a relatively thin bed of brown, nonfossiliferous siltstone at the base of the Ready Pay in three widely scattered localities in the Mimbres region. Four feet of siltstone is found at the base of the Percha shale on North Percha Creek, near an abandoned Forest Ranger station. A sample of the silt was thin-sectioned (Fig. 7, *d*); and, when compared with the thin section from the "Spirifer zone" of the Sly Gap, very few differences were noted except for grain size. A similar occurrence was studied at the base of the Percha shale in Bear Mountain, where approximately 6 feet of brown siltstone is found. Bowsher[63] has also reported approximately 4 feet of brown siltstone below the Ready Pay member in Decker Draw, due east of P. A. Mountain, in the Hillsboro quadrangle.

Until a fauna is found in the Ready Pay or the brown siltstone underlying the Percha shale, or a new locality reveals the actual succession of formations, equivalency of the black shales will be difficult to establish. The brown siltstones can conceivably be either Canutillo, Onate, or Sly Gap in age, although the latter age is most probable. Possibly

[63] Personal communication.

the final solution of this problem will be through the use of well-log data. In the Rattlesnake Field in northwest New Mexico a deep test well penetrated 140 feet of green shale, dense tan limestone, and dense brown dolomite, believed to be Devonian in age.[64] In the event of deep wells being drilled in southern New Mexico, well-log data will unquestionably permit more positive surface identifications, as well as correlations with subsurface strata in areas outside of southern New Mexico.

The Percha shale fauna was first studied by E. M. Kindle;[65] and M. A. Fritz[66] has recently described eight new species of bryozoans from the Box member, which is undoubtedly the age equivalent of the Ouray limestone of Colorado. Collections from the Box member of the Percha shale made by the writer indicate that approximately 80 per cent of the fauna has been described. Revision of the described forms and description of the remaining unidentified fossils will be undertaken sometime in the future.

CONCLUSIONS

There are many lines of evidence which must be more thoroughly explored before a completely satisfactory picture of the Devonian of New Mexico can be presented.

The Onate formation is difficult to interpret. The lack of a diagnostic faunal assemblage at present precludes definite correlations with formations to the east or west. The exact age of the Onate may finally fall somewhere within the time range from the Onondagan to the late Hamilton.

A rather complete picture of the Sly Gap formation is presented in southern New Mexico, but its relationships with the Percha shales to the west are unknown. On the basis of faunal evidence, the Sly Gap is definitely distinct from the Percha fauna and that of the Ouray limestone. Only long-ranging genera and species have been found both in the Percha and Sly Gap assemblages, and only a relatively low percentage of these. Through the study of large Devonian collections in Walker Museum, it was found that more Sly Gap forms were identical with species from the Hackberry of Iowa than with those from any other formation. There are several differences between the two faunas, and the Sly Gap is probably older than the Hackberry. The Sly Gap fauna is probably younger than the fauna of the Independence shale of Iowa, inasmuch as several forms that appear in the Sly Gap are only found stratigraphically higher than the Independence. If these conclusions concerning the Sly Gap are correct, it is older than the Martin limestone of Arizona, which is definitely a time equivalent of the Hackberry. According to A. A. Stoyanow,[67] the stratigraphic position of the Martin limestone is lower than the Lower Ouray formation on Pinal Creek, north of Globe, Arizona. The "Lower Ouray" designation has been employed as a working term by Stoyanow to indicate sediments containing the fauna found in the Ouray limestone of Colorado, which is undoubtedly an equivalent of the Box member of the Percha shale. Although there is no known sec-

[64] "Rattlesnake Field: The Oil and Gas Resources of New Mexico," *N.M. School of Mines, State Bur. Min. and Min. Res. Bull. No. 18* (2d ed., 1942), p. 119.

[65] Ftn. 3 (1909).

[66] Pp. 31–41 of ftn. 19 (1944).

[67] Pp. 489–93 of ftn. 7 (1936).

tion exposing the Sly Gap below the Percha shale, any correlation of the New Mexico formations with those of Arizona brings out the fact that the Sly Gap formation is older than the Percha.

Devonian specimens figured by C. W. Merriam,[68] from the "Cyrtospirifer zone" of the Devils Gate formation, Nevada, resemble many forms in the Sly Gap fauna; and a correlation of time-stratigraphic equivalency of the two formations is likely.

The geologic history of southern New Mexico during late Devonian and Mississippian times seems to indicate low, unstable land masses and fluctuating seas. There is the possibility that the Sly Gap formation may be represented by three facies: (1) the lower fossiliferous, typical Sly Gap sediments; (2) the Contadero formation as a lentil; and (3) the Ready-Pay-member type of strata, possibly representing deposition in a low marshy area on the fringes of the main Sly Gap sea (Fig. 10). Sedimentation under such conditions would in part explain the failure to find the Sly Gap fauna in the area in which the Box member of the Percha shale occurs.

[68] Ftn. 67.

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ORIENTATION ANALYSIS OF FINE-GRAINED CLASTIC SEDIMENTS: A REPORT OF PROGRESS

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ABSTRACT

The purpose of this investigation is to extend methods of orientation analysis to fine-grained clastic sediments in order to determine the character of their last depositional environment, including the nature and direction of movement of the depositing agent. A two-dimensional method, based upon statistical analysis of measurements of the projections of grains on the bedding plane, is developed. Two measurements are made on each grain projection: (1) elongation azimuth and (2) end position (defined as the position of the larger end of the grain with respect to the smaller end). Grains from laboratory fluvial and eolian environments exhibit a pronounced preferred elongation parallel to the direction of flow of the depositing agent and a marked tendency to lie with their larger ends upcurrent. Results of this study indicate that the source direction of fine-grained fluvial and eolian sediments can be determined uniquely by use of the two measurements and suggest that the agent of deposition of a sediment may be determinable from its orientation pattern.

INTRODUCTION

Investigators concerned with the primary structure of sedimentary rocks must attach certain significance to the mineral composition, shape, surface texture, size, and arrangement of the component grains. The degree that these physical properties characterize the environmental history of a sediment, however, is a matter of considerable uncertainty, arising largely from an inability to evaluate the observed data properly. Interpretation is made difficult by the complexity of these properties, which results from differences in the number, the succession, and the nature of the environments to which each mineral grain may be subjected. Collectively, the visible effects of the environments too often tend to confuse rather than to facilitate reconstruction of the depositional history of a sediment, and in such instances it becomes necessary to interpret the last phase of the history on the basis of certain selected physical properties. Of these, the arrangement of mineral grains in unmetamorphosed sediments appears to be least influenced by environments previous to the last and hence should provide one of the most reliable bases for interpretation of the ultimate depositional environment of a sediment.

The existence of preferred dimensional orientation of pebbles in coarse-grained unconsolidated deposits has been demonstrated by Hakon Wadell,[1] Konrad Richter,[2] and W. C. Krumbein.[3] It is reasonable to expect, therefore, that orientation patterns exist in fine-grained clastic materials. If such patterns remain fixed in a sediment after deposition and show tendencies to be diagnostic of particular environments of deposition, they should provide a means of deter-

[1] "Volume, Shape and Shape-Position of Rock Fragments in Open-Work Gravel," *Geografiska Ann.*, Band XVIII (1936), pp. 74–92.

[2] "Die Bewegungsrichtung des Inlandeises, rekonstruiert aus den Kritzen und Längsachsen der Geschiebe," *Zeitschr. f. Geschiebeforschung*, Band VIII (1932), pp. 62–66; "Gefüge und Zusammensetzung des Norddeutschen Jungmoränengebietes," *Greifswald Univ., Geol.-Pal. Inst. Abhand.*, Heft XI (1933), pp. 1–63.

[3] "Preferred Orientation of Pebbles in Sedimentary Deposits," *Jour. Geol.*, Vol. XLVII (1939), pp. 673–706; "Flood Deposits of Arroyo Seco, Los Angeles County, Calif.," *Bull. Geol. Soc. Amer.*, Vol. LIII (1942), pp. 1386–91.

mining the conditions under which a sediment was last deposited.

Wadell[4] and Krumbein[5] have developed methods by which the position in space of an individual pebble may be defined. By these methods data are accumulated on the positions in space of many coarse particles and analyzed statistically to determine the existence of preferred dimensional orientations. However, application of the procedure of Wadell and Krumbein is limited by difficulties encountered when the particles are too small to handle and measure individually.

It is the purpose of this investigation to develop a method of orientation analysis for fine-grained particulate materials, to investigate the existence of diagnostic orientation patterns in fine-grained sediments, and to establish principles concerning such patterns. If the quantitatively predominant fine-grained clastic sediments can be analyzed in this manner, the results obtained should prove valuable in the solution of such problems as (1) determination of the direction of the source of a sediment, (2) determination of the agent of deposition of a sediment, and (3) correlation among sedimentary formations.

At present the authors wish to report a method of determining the orientation of the coarser elements of fine-grained particulate materials. Preliminary results employing this method reveal the presence of what may prove to be diagnostic orientation patterns in fine-grained sediments; at least such patterns were discovered in the laboratory sediments examined. Because of the restricted conditions under which the sediments were deposited in the laboratory, the results are limited in scope, and conclusions of a general nature must await more intensive and extensive research.

[4] Ftn. 1 (1936).

[5] Ftn. 3 (1939).

TERMINOLOGY

As used herein, the term "preferred orientation" refers to a spatial arrangement of grains which can be shown by accepted statistical methods to exhibit a significant deviation from random disposition (i. e., from disposition with an equal number of grains in each possible position). Orientation is considered only from a geometrical standpoint—the dimensional orientation of E. B. Knopf and Earl Ingerson.[6] Since no crystallographic directions are employed, the term "fabric," used by these authors to signify "not only the pattern produced by the external shape of rock constituents, but also the pattern of the internal elements down to the very space lattice of the rock minerals,"[7] is not applicable. Consequently, the term "dimensional orientation pattern," or simply "orientation pattern," understood to refer to the summation of the dimensional orientations of the component grains of a material, is used throughout this discussion. Data concerning dimensional orientation patterns would have to be combined with all other structural and textural data to describe the fabric of any material; such an integration is beyond the scope of this paper.

New terms used in this study are shown by italics.

CLASSIFICATION OF DIMENSIONAL ORIENTATION PATTERNS

To facilitate discussion of experimental data, it is desirable to establish a terminology by which the various types of dimensional orientation patterns may be

[6] "Structural Petrology," *Geol. Soc. Amer., Mem. VI* (1938), p. 10.

[7] *Ibid.*, p. 12.

described. Classification by agent is immediately suggested and affords such descriptive terms as "eolian," "aqueous" (fluvial, littoral, marine, lacustrine), and "gravity" or "angle of repose" (subaerial, subaqueous) dimensional orientation patterns. In addition, division of each of these agent *classes* into the following *groups* is helpful.

1. *Translational orientation pattern.*—Dimensional orientation pattern of material while it is in the transporting medium and before deposition.[8] This group may be subdivided into (*a*) suspension, (*b*) saltation, and (*c*) traction *phases* or into (*a*) bed-load and (*b*) suspended-load phases.[9]

2. *Depositional orientation pattern.*—Dimensional orientation pattern resulting from deposition and continued action of the transporting agent after translatory motion of the particles has ceased. This pattern must be developed during deposition and in the period when the deposit is in contact with the transporting medium. It may vary with the length of time between deposition and burial.

3. *Secondary orientation pattern.*—Dimensional orientation pattern resulting from any factors (natural or artificial) other than the agent of deposition. Phases of this group would include patterns due to weathering, metamorphism, effects of organisms, etc.

The above method of classification has the advantage of being complete and of using terms that are concise and suggestive of the nature of the orientation pattern. Orientation patterns may now be described uniquely by the use of combinations of the terms suggested—for example, translational aqueous orientation pattern, translational fluvial orientation pattern (saltation phase), depositional marine orientation pattern, and secondary recrystallization orientation pattern. A more complicated terminology may be required to describe results of advanced research, but this system is adequate for present purposes.

METHOD OF INVESTIGATION

The complete analysis of the position of a particle in a stratum is a three-dimensional problem and should include the determination of both azimuth and angle of dip of the grains. In describing the orientation of coarse clastic sediments, Krumbein[10] presents methods which satisfy these requirements. With fine-grained sediments, however, manipulation of the particles becomes difficult, precluding the application of Krumbein's methods. No satisfactory means of handling the small grains individually and no other successful three-dimensional method of analysis was discovered in this investigation. Consequently, the writers have investigated the possibilities of using a less complete system of two-dimensional analysis which does not involve removal of the particles from their original position. This method, depending primarily on the outline shape of the projection of grains, apparently provides sufficient diagnostic data to merit use.

A specimen of the material whose orientation pattern is to be investigated is photographed through a standard petrographic microscope. Magnification is increased further by photographic enlargement until the dimensions of the grains

[8] The translational behavior of nonspherical particles has been investigated by Krumbein, "Settling-Velocity and Flume-Behavior of Nonspherical Particles" *Trans. Amer. Geophys. Union, 1942*, Part II (1942), pp. 621–33; see also Wadell, "Some New Sedimentation Formulas," *Jour. App. Phys.*, Vol. V (1934), pp. 281–91.

[9] H. A. Einstein, A. G. Anderson, and J. W. Johnson, "A Distinction between Bed-Load and Suspended-Load in Natural Streams," *Trans. Amer. Geophys. Union, 1940*, Part II (1940), pp. 628–33.

[10] Ftn. 3 (1939).

are raised to lengths ranging from $\frac{1}{2}$ to 1 inch. Best results are obtained by using low-power magnification with the microscope and photographing onto a $3\frac{1}{4}'' \times 4\frac{1}{4}''$ negative. In this manner a large number of grains in sharp focus is obtained, and magnification is achieved principally through enlargement of the negative. Correct orientation of each photograph is insured by suspending a needle point or finely drawn glass capillary over the northeast quadrant of the microscope field.[11]

The method of determining and statistically studying the grain orientation is based upon the projection of the grains visible on the stratification plane. The bedding plane was selected as a surface of prime importance for several reasons: (1) it is generally recognized easily in outcropping strata and in drill cores; (2) knowledge of the orientation of the grains in the stratification-plane direction is necessary to give the source direction of the sediment; (3) the surface approaches a plane; (4) slight weathering of the bedding surface commonly results in the grains' standing out in relief; hence, in the photograph the grain boundaries are distinct.

The authors are of the opinion that it may be possible to determine the true orientation of a grain from its projections on three mutually perpendicular planes. However, calculations based upon this assumption, considering the grains to be triaxial ellipsoids, have yielded results expressible only in mathematical terms of an order too high to permit practical usage; whereas the two-dimensional method, using the projection of grains on the bedding surface, has as its greatest advantage the simplicity of calculation and, as far as the investigation has been carried, does reveal the nature of dimensional orientation patterns.

A practical approach to three-dimensional analysis would be to use projections on two additional planes, one parallel and one perpendicular to the direction of movement of the grains at the time of deposition (as determined by analysis of the bedding-plane orientation pattern) and both perpendicular to the bedding plane. This method should disclose imbricate or edgewise structures and, being more complete, probably would permit reliable interpretation of depositional environments in some cases where bedding-plane analysis alone would be inadequate. The study of this method should await development of the bedding-plane method of determining current direction, upon which it depends, and which in itself yields one of the major results desired. It is the authors' intention to investigate this more nearly complete three-dimensional method at a later date.

The nature of an orientation pattern is revealed only by a statistical analysis of the attitude of many grains. Study of the positions of individual grains is of little significance in a system so complicated by irregular grain shapes and wide range of orientation. For the collective study necessary in such an analysis, statistical methods provide means of synthesizing the data, and criteria for judging whether the observed orientation distributions are due to chance alone or follow some other law of distribution (i.e., whether they exhibit random or preferred orientation).[12]

[11] With further projection-orientation studies, greater rapidity of analysis with satisfactory results may be obtainable by use of a microprojector. See W. D. Pye, "Rapid Methods of Making Sedimentational Analyses of Arenaceous Sediments," *Jour. Sed. Pet.*, Vol. XIII (1943), pp. 86–88.

[12] The statistical notation used herein is that of J. F. Kenny, *Mathematics of Statistics*, Parts I and II (New York: D. Van Nostrand Co., 1939).

PROBABILITY THEORY

In this investigation it has been necessary to determine whether the distributions obtained are due to natural factors or represent merely chance deviations from random distributions. For this purpose the theory of the sampling of attributes has been employed.[13] By this method it is possible, for example, to determine the probability of occurrence of sixty heads and forty tails in throwing a hundred coins. If the calculated probability in such cases is less than 0.01 (1 chance in 100), a conservative limit, the deviation from chance is said to be significant, and it may be concluded that the coins are biased. On the other hand, if the probability is greater than 0.05 (1 chance in 20), the deviation from the expected value of fifty heads and fifty tails is not regarded as significant and may be accepted as due to chance fluctuations. It should be noted that the 0.01 and 0.05 limits are arbitrary values that have been adopted by statisticians as a result of experience; in the range 0.01–0.05 no conclusions can be drawn with assurance. Tests according to the sampling-of-attributes theory may be made to evaluate the significance of the deviation of any variable from its expected value.

Each distribution in this study has been tested to determine whether or not it exhibits a significant deviation from the value expected by chance alone. If the calculation indicates that the deviation is significant, it is concluded that the factor other than chance is a physical process of sedimentation.

According to the sampling-of-attributes theory, if an event has the probability of occurence p, the probability that the number of occurrences x in s trials will differ from the expected number $E = sp$ by as much as d is given approximately by $P_\delta = 1 - Q_\delta$, where

$$Q_\delta = 2\int_0^\delta \phi(t)\,dt,$$

$$\delta = \frac{d - \frac{1}{2}}{\sigma},$$

$\phi(t) = (2\pi)^{-\frac{1}{2}} e^{-\frac{t^2}{2}}$, the equation of the normal probability curve,

$\sigma = (spq)^{\frac{1}{2}}$, the standard deviation,

$q = 1 - p$, the probability of non-occurrence.

It will be noted that the integral in the equation for Q is that evaluated in mathematical tables as "area under the normal probability curve." The probability calculations used in this study are all made according to these equations.

TYPES OF MEASUREMENTS

Two main types of measurements are employed in this analysis: (1) elongation direction (p. 251) and (2) end position (p. 251) with respect to that direction. Both elongation direction and end position are included in the concept of preferred dimensional orientation, and, as demonstrated below, a combination of these two dimensions is necessary to determine uniquely the direction of motion of the depositing agent of a sediment. Both of these dimensions might well be included in Wadell's[14] term "shape-position."

ELONGATION DIRECTION

Three methods of measuring the elongation direction of grain projections, each with accompanying advantages and disadvantages, are recognized in the present study.

[13] *Ibid.*, Part II, pp. 24–26.

[14] Ftn. 1 (1936).

1. *Long-dimension elongation.*—Direction of elongation is considered as that of the longest line which can be drawn on the grain projection (Fig. 1). This dimension is the projection equivalent of the three-dimensional measure of elongation used by Krumbein[15] for coarse clastic sediments. It has the advantage of ease of determination but loses value by its susceptibility to irregularities of the grain projections.

2. *Least-projection elongation.*—Elongation direction is considered as that of the two parallel lines with the minimum amount of separation which can be drawn tangent to the grain projection (Fig. 1). In practice these lines are constructed readily with the aid of a parallel rule. This dimension has theoretical importance in that the least projection in most cases probably offers least resistance to a circumfluent medium.

3. *Center-of-area elongation.*—Elongation direction is considered as that of the longest straight line which can be drawn through the center of area of the grain projection. (The term *center of area* is defined herein as the point of intersection of all lines bisecting the area of a grain projection. It is, therefore, the two-dimensional equivalent of center of mass.) This method has the disadvantage of requiring the use of a planimeter on each grain projection. However, it probably enjoys the soundest theoretical basis of the three methods, since the center of area of a grain projection is in most cases nearly above the center of mass of the grain and, hence, above the point about which the grain pivots when suspended in a fluid.

END POSITION

Knowledge of preferred elongation direction, and the fact, as demonstrated below, that this direction is parallel to the current direction in fluvial and eolian environments, limits the current to two possible directions. For example, if the elongation trend of grains in a fluvial environment is found to be statistically north-south, the current could have flowed either from the north or from the south. A method of making the correct choice between the two possibilities is provided by determining the position of

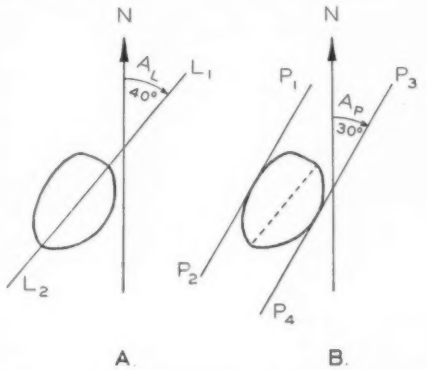

FIG. 1.—Comparison of long dimension and least projection measures of elongation azimuth. *A*, Grain with long dimension axis (L_1L_2) making an azimuth angle (A_L) of 40° from north (N). *B*, Same grain with least projection tangents (P_1P_2, P_3P_4) showing a least projection azimuth (A_P) of 30°.

the larger end of the grain with respect to the smaller end, herein defined as *end position*, for each grain, in terms of the two opposite directions indicated by the elongation analysis (north and south in the example just cited). Use of this measurement rests on the assumption that the larger end of a grain tends to be upcurrent from the smaller end; a relation known to exist in falling drops of water. In this position, corresponding to the so-called "tear-drop" streamlined shape, resistance to flow is at a minimum.

End position, as defined, is a three-dimensional measurement—a compari-

[15] P. 677 of ftn. 3 (1939).

son of the relative volumes of the two parts of a grain delimited by the plane bisecting its elongation axis, and perpendicular to that axis. As such, it may not be the exact property desired; in fact, even the least-resistance principle hypothesized may not be valid. However, the results obtained by use of one of the following measurements, each of which represents a projection approach to the determination of the three-dimensional quantity, make use of the concept of end position valuable.

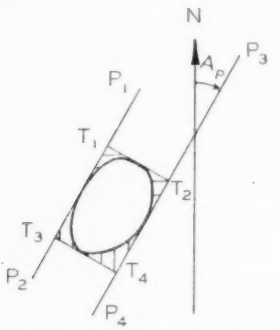

FIG. 2.—Least projection rectangle ($T_1T_2T_3T_4$) constructed about Figure 1 grain projection by erecting tangents (T_1T_2, T_3T_4) perpendicular to the least projection tangents (P_1P_2, P_3P_4). The smaller area (horizontally ruled) outside of the grain and within the least projection rectangle is at the north (N) end of the rectangle. Hence the greater area is in the north portion of the grain—i.e., the area direction is north.

Three methods of describing end position, one corresponding to each of the elongation measurements, afford possible quantitative means of distinguishing the larger end of a grain from its two-dimensional projection.

1. Intersection of axes position: In conjunction with the long-dimension axis, and corresponding essentially to the intermediate axis of Krumbein,[16] is the longest line that can be drawn on a grain projection perpendicular to its long-dimension direction. If it is not affected too much by irregularities in the grain outline, the position of the intersection of this line on the long-dimension axis, with respect to the center of that axis, should in most instances be indicative of the end position of the grain.

2. Area direction: The position of the greater area of a grain projection with respect to the center of its elongation axis, herein defined as *area direction*, is evidently the projection equivalent of end position. The precise method of determining this dimension is by using a planimeter to measure the area on either side of the perpendicular bisector of the elongation axis. As such a process is somewhat laborious, the authors have devised a method, requiring only visual inspection, which provides a reliable measure of end position. In this method the *least-projection rectangle* (rectangle formed by the least-projection tangents to the grain projection and the perpendicular pair of tangents) is constructed around each grain. The end of the grain having the greater area may then be determined by visual estimation of the half of the least-projection rectangle having the least area outside of the grain projection (Fig. 2). With a little practice the area direction of most grains may be determined with confidence by this method. Grains to which the visual method cannot be applied with certainty may be omitted without harm, the omission having the effect of reducing the importance, in the total sample, of minor irregularities on nearly symmetrical grains.

3. Center-of-area position: The position of the center of area of a grain with respect to the midpoint of the center-of-area elongation line is a measure of the grain's end position. As previously stated, the procedure for determining the center of area is lengthy, and, hence,

[16] *Ibid.*

determination of end position by this method is not accomplished with the rapidity of methods (1) and (2).

METHOD EMPLOYED

The measures of elongation, coupled with the corresponding means of determining end position, give three complete methods of analysis: (1) long-dimension elongation—intersection of axis position; (2) least-projection elongation—area direction; and (3) center-of-area elongation—center-of-area position. It is significant that both the theoretical soundness and the practical difficulty increase from method (1) to method (3). In view of this fact the authors have selected method (2) as the one which embodies the optimum combination of theoretical accuracy and operational facility. In the case of method (1) an orientation of the dimensional elongation of the grain projection is revealed which closely approaches the azimuth of current direction. On the other hand, this method fails to demonstrate that a relationship exists between the end position and the direction of current flow of the transporting medium (p. 255).

This weakness is overcome by method (2), which not only reveals a more decided preferred elongation (p. 255) but also gives a diagnostic measure of end position (p. 255). As method (2) yields satisfactory results, no study of the more difficult method (3) has yet been undertaken. It is possible, however, that depositional orientation patterns for sediments deposited in environments not investigated in this study may be revealed by method (3) and not by method (2). Until such proves to be the case, the advantages of ease and speed of determination of the least-projection–area-direction method should not be sacrificed.

ANALYSIS OF LABORATORY FLUVIAL SAND

The concepts discussed above arose from, or are based upon, an investigation conducted (1) to determine whether orientation patterns could be discovered in sediments deposited in the laboratory and, if so, (2) to determine the relationship between the pattern and the direction of the current which carried the component particles. The investigation included a study of the orientation of sand grains deposited under conditions which would closely simulate those in fluvial, eolian, and subaerial gravity (angle of repose) environments.

The apparatus used to produce the fluvial environment was a stream table 4 feet long, $3\frac{1}{2}$ inches wide, and $1\frac{1}{4}$ inches deep, constructed in a series of removable sections. Sand deposited on any part of the table could in this way be removed and studied without disturbing the grains. From a well at one end of the table, water was made to flow as a sheet slightly more than $\frac{1}{2}$ inch thick, covering the entire width of the trough. The inclination of the table was equivalent to a gradient of 90 feet per mile. Unsieved sand from the Lake Michigan beach along the Northwestern University campus[17] was poured into the upper end of the trough at a rate sufficient to maintain a small ridge of sand over which the water passed and gathered its load. Despite the uniform flow from the well, several rather strong cross-currents developed and could not be eliminated. Arbitrarily designating the source direction as due north, the cross-currents flowed

[17] Median diameter 1/10 mm. For a mechanical analysis of this sand see J. P. Todd, "Survey of the Near-Shore Deposits of Lake Michigan Adjacent to Northwestern University" (unpublished Master's thesis, Northwestern University, 1937), p. 32.

from northeast to southwest and from northwest to southeast. After ½ inch of sand had been deposited in the north (upstream) end of the trough, the flow of water was stopped and the sand allowed to dry. The section of the table from 4 to 16 inches below the sand source was then removed and placed under a petrographic microscope, the axis of the trough (north-south direction) being kept parallel to one of the cross-hairs. Twelve photographs, beginning at a point 8 inches below the sand source and continuing down the axis at ½-inch intervals, were taken through the microscope. It should be noted that the photographs constitute grain projections in a direction equivalent to the stratification plane. The first four pictures were of the upstream portion that apparently was unaffected by cross-currents, whereas the remaining ones were in the cross-current area. All the grains appearing distinctly on the enlarged pictures were then outlined in pencil, the pictures and grains numbered, and north lines drawn on each picture. On each of 507 grain projections thus delimited, the following measurements were made:

1. Azimuth angle of long-dimension axis: The long-dimension axis was first drawn on each grain projection, its position being determined accurately by the use of dividers. The azimuth of this axis, measured clockwise 0°–180° from north was then determined to the nearest degree for each grain. This is a measure of long-dimension elongation.

2. Azimuth angle of least-projection axis: Least-projection tangents were constructed about each grain with the aid of a parallel rule. Azimuth angles were measured in the same manner as those of the long-dimension axis and recorded as a measure of least-projection elongation.

3. Position of intersection of axes: The longest possible line perpendicular to the long-dimension axis was constructed on each grain projection. The position of the intersection of this line with the long-dimension axis was then determined as north or south of the midpoint of the axis. This position was recorded as either north or south for all grains having elongation azimuths within 89.5° of the current direction.

4. Area direction: Least-projection rectangles were completed about each grain and the area directions determined visually by estimating the half of the rectangle having the least area outside the grain. The area directions of 447 of the 507 grains were determined, the remaining grains being too nearly symmetrical for reliable estimation by this method.

SUMMARY OF FLUVIAL ELONGATION AZIMUTH DATA

Compilation of azimuth data reveals a pronounced preferred elongation parallel to the direction of movement of the depositing agent. Azimuth distribution histograms of both long-dimension and least-projection directions (Figs. 3 and 4) show a distinct mode of azimuths within 10° of the current direction. In the least projection distribution the 0°–9° and 170°–179° classes (those within 10° of the current direction) each contain 51 grains, an excess of 23 over the expected number (507/18) if the grains were oriented at random. The probability of obtaining such a concentration by chance alone is 0.00000 (less than 1 in 200,000), a value far less than the 0.01 accepted by statisticians to measure the greatest allowable deviation from any acceptable theory of distribution.[18] These

[18] The 0.00000 probability is obtained by the method described on pp. 250–54, the actual cal-

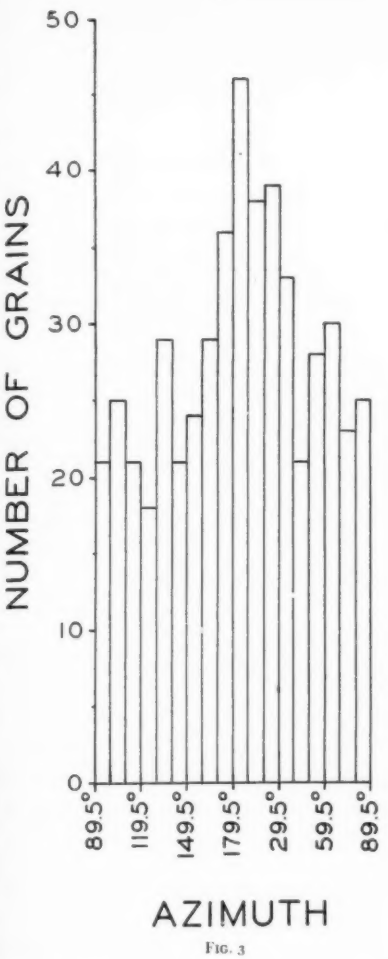

FIG. 3

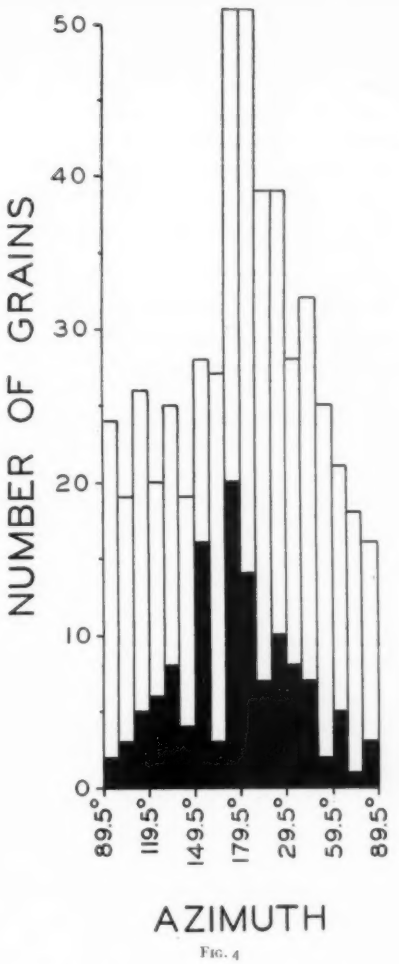

FIG. 4

FIG. 3.—(*Left*) Histogram of long dimension azimuth distribution of laboratory fluvial sand. Current direction = 0.0° (=180.0°). (Class boundaries, 89.5°, 99.5°, etc., are shown for classes 90°–99°, 100°–109°, etc.)

FIG. 4.—(*Right*) Histogram of least projection azimuth distribution of laboratory fluvial sand, showing distribution of grains in area with cross-currents (*white*) and in area without visible cross-currents (*black*). Current direction at source =0.0° (= 180.0°).

data constitute unquestionable evidence that grain elongation is not random but controlled by the depositing agent.

A statistical study of the two distributions of azimuths measured 0°–180° from the known current direction yields the characterizing constants shown in Table 1. It is evident that the current direction indicated by the mean of the azimuths in either type of measurement differs from the actual current direction (0.0°) by an insignificant amount. The skewness constants indicate a slightly greater number of azimuths in the 90°–180° than the 0°–90° range, a feature attributed to the effect of the irregularity of the strong cross-currents. The fact that the kurtosis values are less than 3.0 indicates that the distribution curves are slightly "flatter" than the normal probability curve.

Comparison of the long-dimension and least-projection elongation distributions gives a slight advantage to the latter. The long-dimension modal frequency of 46 (Fig. 3) is less distinct than the two classes containing 51 least-projection azimuths. In addition, the long-dimension azimuth distribution exhibits a greater standard deviation, a greater skewness, and a mean differing by a larger amount from the true current direction than the corresponding least-projection values. The least-projection method is, therefore, slightly more critical than the long-dimension method—a factor which might prove decisive in an investigation of specimens from environments productive of more nearly homogeneous orientation patterns.

TABLE 1

CHARACTERIZING CONSTANTS OF DISTRIBUTIONS OF AZIMUTHS MEASURED 0°–180° FROM KNOWN CURRENT DIRECTION FOR SAND DEPOSITED BY RUNNING WATER (FLUVIAL DEPOSITIONAL ORIENTATION PATTERN)

	Long-Dimensional Azimuthal Direction (Fig. 3)	Least-Projection Azimuthal Direction (Fig. 4)
Mean	3.3°	0.0°
Standard deviation	48.1°	45.8°
Skewness	− 0.136	− 0.124
Kurtosis	2.22	2.17

In order to determine the effect of cross-currents on elongation azimuth distributions, the total sample was divided into two portions, the first containing 124 grains from the region in which no cross-currents were observed, the second comprising the 383 grains from the area in which cross-currents were noted (Fig. 4). Table 2 is a summary of the azimuthal distribution of the least-projection tangents. It is apparent that cross-currents have the ef-

culation being given here as an example of the method used throughout this paper.

Since, by chance alone there are 18 equally likely and mutually exclusive azimuth classes, the probability p of occurrence of azimuths in any class is 1/18.

$s = 507$ total grains counted $\quad p = 1/18$

$q = 1 - p = \frac{17}{18}$ $\quad x = 51$ grains in classes with elongation azimuths within 10° of current direction

$E = sp = \frac{507}{18} = 28$ = expected number of grains in any 10° class

$d = x - E = 51 - 28 = 23$

$\sigma = (spq)^{\frac{1}{2}} = \left(\frac{507 \cdot 17}{18 \cdot 18}\right)^{\frac{1}{2}} = 5.16$

$\delta = \frac{d - \frac{1}{2}}{\sigma} = \frac{23.0 - 0.5}{5.16} = 4.37$

$P_\delta = 1 - 2\int_0^{4.37} \phi(t)dt = 1 - 2(0.50000)$

= 0.00000 = probability of occurrence of a deviation from the expected value, in *any* class, of as much as 23. Hence, the probability of occurrence of this modal value in one of the two classes adjacent to the azimuth of current direction is $2/18 \cdot 0.00000 = 0.00000$.

fect of increasing the standard deviation, increasing the skewness, and decreasing the kurtosis of the distribution. The fact that the mean azimuth of the portion affected by cross-currents differs from the true current direction by only 0.9° suggests the applicability of this system of analysis to materials from natural environments in which cross-currents are common.

Insufficient data are available from this investigation to determine whether the mean or modal azimuth of elongation gives the more accurate portrayal of current direction. The distributions summarized above are so symmetrical that their mean and modal values nearly coincide. It is suggested, however, that the mean is affected by cross-currents to a greater degree than is the mode and, hence, that the mode is the more dependable value.

TABLE 2

CHARACTERIZING CONSTANTS OF DISTRIBUTIONS OF LEAST-PROJECTION AZIMUTHS MEASURED 0°–180° FROM KNOWN CURRENT DIRECTION FOR SAND DEPOSITED BY RUNNING WATER (FLUVIAL DEPOSITIONAL ORIENTATION PATTERN)

	Distribution Where Cross-Currents Are Absent	Distribution Where Cross-Currents Are Present
Mean	177.0° (−3.0°)	0.9°
Standard deviation	39.0°	47.7°
Skewness	0.073	−0.194
Kurtosis	2.54	2.09

SUMMARY OF FLUVIAL END-POSITION DATA

The arrangement of grains with their larger ends statistically toward the source of material is demonstrated conclusively by the measurement of area direction. Of the 447 grains amenable to estimation by the visual method, 62 per cent had their greater area toward the current. Since the probability of such a deviation from the expected random value of 50 per cent in a sample of this size is 0.00000,[19] there can be no doubt that the observed preponderance of area direction in the upcurrent position is due to physical factors and does not represent merely chance deviation from a random distribution. The accepted limit of probability of 0.01 as indicative of a preferred distribution is reached by every series of 200 consecutive grains in the sample; hence, area direction data for 200 grains probably could be considered as sufficient to yield reliable results in this type of distribution.[20]

The measurement of end position in the investigation using the method of position of intersection of axes (p. 251) does not indicate the presence of a preferred position of the large ends of the grains. Approximately 52 per cent of the intersections fall on the north side of the midpoint of the long-dimension axis. The probability of obtaining this deviation of 2 per cent from the expected value of 50 per cent, if the distribution were due to chance alone, is 0.467 (nearly 1 in 2). Hence this method is either entirely unsatisfactory or requires measurement of a much larger number of grains than does the method of area direction. Its weakness is attributed to the important effect of slight irregularities of grain outline upon the location of the two axes.

[19] The 0.00000 probability is calculated according to the method illustrated on p. 252.

[20] In order to determine the least number of consecutive grains in the observed distribution required to have a probability of occurrence by chance below the 0.01 significance level, several trial probabilities were calculated. Each of these included data from photographs containing the least excess over the value expected by chance. It was found that the least number of consecutive grains required for a probability of less than 0.01 included 206 grains.

ANALYSIS OF LABORATORY EOLIAN SAND

The problem of determining whether depositional patterns similar to those produced by running water result from wind action was investigated in the laboratory by blowing sand in one direction. In order to produce wind blowing with constant force and limited direction at the source, a current of air from a $\frac{1}{8}$-inch hose was directed downward onto the stream table from a point about $\frac{1}{8}$-inch above the floor of the table. Dry sand from the beach of Lake Michigan, as in the fluvial experiments, was supplied at the air source until a layer of sand had been deposited over the entire table. In the process of deposition many grains moved through the air directly to their place of rest, whereas others moved by jumps similar to the saltation movement of grains in a fluvial environment. A section of the stream table was then removed and the surface of the deposit photographed, following the method previously described. In this manner the projection outlines of 642 grains which had been deposited 2 feet from the sand source were obtained. Measurements of (1) least-projection azimuth and (2) area direction were made on these grains.

SUMMARY OF EOLIAN LEAST-PROJECTION DATA

Data from 642 grains show that for least-projection azimuths a mode of 49 exists in the class adjacent to the wind direction and a secondary mode of nearly equal prominence (48) exists in the 160°–169° class (11°–20° from the current direction) (Fig. 5). The excess of 13.3 (49.0 − 35.7 = 13.3) over the expected value of 35.7 has a probability 0.027 of occurring by chance in any class and the probability of 0.003 of chance occurrence in either class adjacent to the current direction. As this value is below the 0.01 significance level, there can be no doubt that the orientation is due to the operation of physical factors.

The characterizing constants obtained by statistical analysis of the eolian least-projection azimuths are shown in Table 3. Here again the mode and mean differ from the current direction by a small amount. Hence, the principles of interpreting current direction from elongation alignment apply to sand from eolian as well as from fluvial environments. The eolian distribution shows a slightly greater standard deviation and smaller kurtosis parameter than the fluvial pattern.

TABLE 3

CHARACTERIZING CONSTANTS OF DISTRIBUTION OF LEAST-PROJECTION AZIMUTHS MEASURED 0°–180° FROM KNOWN CURRENT DIRECTION FOR SAND DEPOSITED BY WIND OF CONSTANT DIRECTION (EOLIAN DEPOSITIONAL ORIENTATION PATTERN)

Mean	1.3°
Standard deviation	49.8°
Skewness	−0.04
Kurtosis	1.94

Although these differences are too small to be analyzed with certainty, they suggest an expectable tendency toward a lesser degree of preferred orientation resulting from wind action than from water currents.

Comparison of the fluvial and eolian azimuth histograms (Figs. 4 and 5) reveals the presence of a secondary mode normal to the current direction in the eolian distribution, a feature indicating that some of the grains rolled about their long axis. If further study reveals that this secondary mode is characteristic of eolian sediments, a means of distinguishing between sands from eolian and fluvial environments is provided. It should be noted, however, that the azimuth distribution may possibly vary more within a single group of eolian or fluvial sedi-

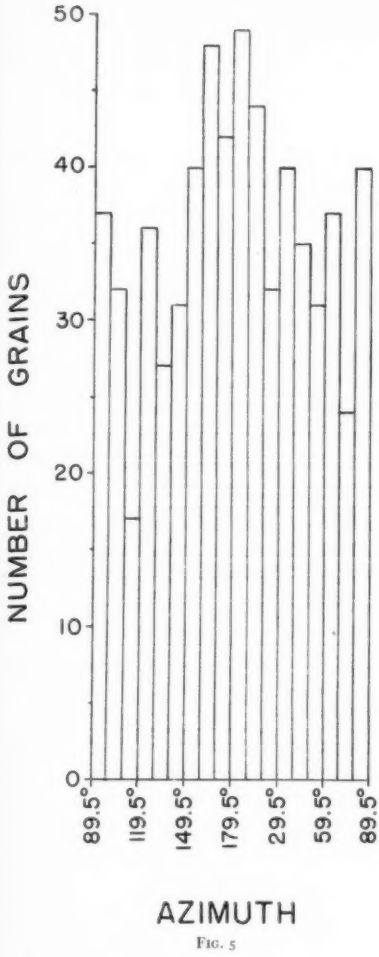

FIG. 5

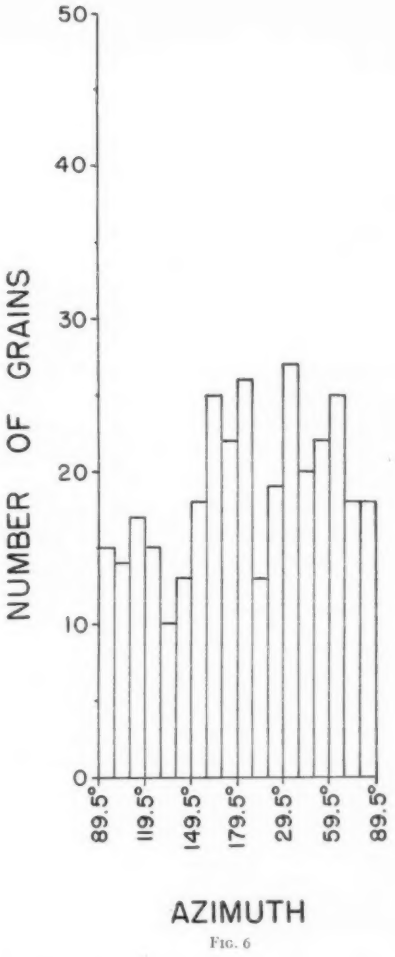

FIG. 6

FIG. 5.—(*Left*) Histogram of least projection azimuth distribution of laboratory eolian sand, showing secondary mode 90° from current direction. Current direction = 0.0° (= 180.0°).

FIG. 6.—(*Right*) Histogram of least projection azimuth distribution of laboratory gravity sand, showing random nature of the distribution. Direction of dip = 0.0° (= 180.0°).

ments, owing to differences in current velocity or other factors not studied, than between these two classes of sediments.

SUMMARY OF EOLIAN AREA DIRECTION DATA

Determination of area direction for the eolian depositional pattern shows that 55 per cent of the grains have their larger ends toward the source. The probability of obtaining this excess over the 50 per cent expected by chance for the 522 grains determined is 0.024. This probability, although greater than the conservative 0.01 limit, is small enough to leave little doubt that the distribution is the effect of the depositing agent and does not represent merely chance deviation from a random distribution.

It is noteworthy that the percentage of grains with area direction toward their source is less than in the fluvial specimens (55 as compared with 62 per cent). Although the actual percentage in each case may be a function of current velocity or some other factor, this difference should be investigated as a possible means of differentiating between sands deposited in fluvial and eolian environments.

ANALYSIS OF LABORATORY GRAVITY SAND

To simulate depositional conditions on the lee side of a dune, dry sand was piled on the sectional stream table and allowed to roll and settle to a position of rest. This was accomplished by inclining the stream table and pouring sand at the top until the angle of repose (32°) was reached and the sand had rolled part way down the trough. Photographs were taken through a binocular microscope set up normal to the surface of the sand, new material being brought into the microscope field by allowing additional sand to roll down from above, until 337 grains had been photographed. The surface of the sand remained at such an angle that grains added at the upper end rolled down either individually or in small landslides.

Measurements were made of (1) least-projection elongation and (2) area direction with respect to the direction of grain movement—that is, the direction of dip of the slope.

The data obtained show that elongation azimuths of grains on a gravity slope are apparently distributed at random (Fig. 6). The mode of elongation direction lies in the class 30°–39° from the direction of dip, and the maximum deviation from the expected value in the grains measured has the probability 0.06 of occurring by chance. Consequently, no interpretation of source direction is possible from such gravity patterns.[21] The lack of preferred elongation is attributed to the small landslides: grains rolling individually down a slope probably tend to have their long axes parallel to the strike of the slope.

Measurements of area direction with respect to the direction of inclination of the slope yield the surprising result that 60 per cent of the grains have their larger ends uphill. The probability of obtaining such a deviation from the expected value of 50 per cent is 0.001. Hence the observed pattern is due to physical causes and does not represent merely chance deviation from a random distribution. It is noteworthy that in this preferred position there is a greater distance between the lower end of the grain and its center of mass than if the larger end of the grain

[21] A method of determining the source direction of material deposited on inclined slopes in dune areas has been described by Parry Reiche, "An Analysis of Cross-lamination: The Coconino Sandstone," *Jour. Geol.*, Vol. XLVI (1938), pp. 905–32.

were downhill. Consequently, more energy is required to roll a grain downhill from this position. Although the absence of preferred grain alignment precludes the application of area-direction measurements to unknown gravity patterns, this energy principle may be of value in future investigations of patterns from other environments.

CONCLUSION

The limited scope of the investigation presented in the preceding pages does not justify the formulation of extensive conclusions concerning dimensional orientation patterns in fine-grained sediments. Nevertheless, it is hoped that certain characteristics which have been observed will serve as a working basis for future study. These observations are summarized as follows:

1. The particles of fine clastic sediments are not oriented at random but exhibit diagnostic patterns which are a product of their depositional environments.

2. A preferred elongation parallel to the current direction exists among grains in fluvial and eolian environments; long-dimension orientation in gravity sands is apparently random.

3. In fluvial, eolian, and gravity environments the larger ends of the grains, as determined by area-direction measurements, tend to lie toward the source.

4. The source direction of fluvial and eolian sediments can be determined accurately by combining measurements of least-projection azimuth and area direction.

5. Differences observed among specimens from the three environments suggest that the agent of deposition of a sediment may be determined from certain characteristics of its dimensional orientation pattern.

ACKNOWLEDGMENTS.—The authors wish to express appreciation to Professor Mason E. Wescott, of the Department of Mathematics, Northwestern University, for his suggestions and constructive criticism of the statistical methods employed herein.

The investigation was conducted in the laboratories of the Department of Geology and Geography, Northwestern University, and this paper, to a large extent, represents the Master's thesis of the junior author.

THE STONES RIVER EQUIVALENTS IN APPALACHIAN REGION[1]

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Virginia Geological Survey, Charlottesville

ABSTRACT

The commonly accepted correlation of the Stones River group of central Tennessee with the Chazy of New York and with beds immediately succeeding the Beekmantown or Knox dolomite in the Appalachian region is no longer tenable. Ulrich's recognition of two Chazyan groups in the Appalachian Valley—the Stones River and overlying Blount—was based upon mistaken identification of formations.

In southwestern Virginia the fauna of the Murfreesboro limestone (basal Stones River) occurs in the Peery limestone about 475 feet above the Beekmantown. The Peery is directly overlain by the Benbolt limestone, which correlates faunally with the Pierce and Ridley formations of the Central Basin of Tennessee. In Bland County, Virginia, Murfreesboro-Ridley equivalents overlie the Athens formation of Ulrich's Blount group. The Lebanon of central Tennessee is linked faunally with the Witten limestone, which corresponds to the beds identified by Butts as Lowville in the Clinch Mountain belt of the Middle Ordovician in Virginia and Tennessee.

Both the stratigraphic position of Stones River equivalents in the Appalachian region and the character of Stones River faunas indicate strongly that the entire Stones River group is post-Chazyan.

ORIGIN AND USAGE OF THE NAME "STONES RIVER"

The name "Stones River group" was originally proposed by J. M. Safford[2] for a succession of limestones exposed along Stone(s) River in the Central Basin of Tennessee. Three subdivisions were named and described: (1) the Stones River beds, at the base, about 75 feet thick; (2) the lower Lebanon limestone; and (3) the upper Lebanon limestone, with a thickness of 110–30 feet. In his next description[3] of this succession the name "Stones River" was replaced by "Trenton or Lebanon formation," and the lithology and fossils of the beds were described for the first time. Five subdivisions were named, in ascending order: (1) Central limestone, about 100 feet thick, forming the "bottom-rock" of the Central Basin and corresponding to the lowest of the three divisions described in 1851; (2) Pierce limestone, composed of thin flaggy beds, with a total thickness of 27 feet; (3) Ridley limestone, thick-bedded and aggregating 95 feet; (4) Glade limestone, light-blue flaggy limestone about 120 feet thick, forming the Cedar Glades of central Tennessee; and (5) the Carters Creek limestone, thick-bedded, dove-gray, totaling 50–100 feet. These units are well displayed in Rutherford County, Tennessee, in the middle of the Central Basin.

In eastern Tennessee, Safford[4] referred to the succession above the Knox and below the Clinch sandstone as the "Trenton and Nashville series," in which he recognized seven divisions. All were given informal lithologic names, except the lowest, which was called the "Maclurea Limestone." Whether Safford then regarded the Trenton group of the Central Basin to be well represented in eastern Tennessee is not clear; but, since the lowest division—the Maclurea limestone—was assigned to the "Trenton and Nashville series," it is only reasonable to infer that Safford regarded the Maclurea limestone to be equivalent to

[1] Published by permission of the state geologist of Virginia.

[2] "The Silurian Basin of Middle Tennessee, with Notices of the Strata Surrounding It," *Amer. Jour. Sci.*, Vol. XII (2d ser., 1851), pp. 352–61.

[3] Safford, *The Geology of Tennessee* (Nashville, 1869), pp. 258–68.

[4] *Ibid.*, p. 228.

part of the "Trenton (Lebanon) formation" of central Tennessee. Safford, according to N. H. Winchell and E. O. Ulrich,[5] abandoned the name "Stones River" in 1869, under the impression that the group was strictly equivalent to the New York Trenton.

J. M. Safford and J. B. Killebrew[6] subsequently referred to the succession as the "Lebanon (Trenton) group." The name "Lenoir" was introduced to replace †Maclurea limestone. It was defined as overlying the Knox and underlying the variegated marbles of the Knoxville region, which were believed to represent the "Lebanon group" of the Central Basin. The Lenoir was described as the correlative of the New York Chazy. Thus Safford and Killebrew in 1876 regarded the entire "Lebanon group" as younger than the Chazy.

Winchell and Ulrich[7] later revived the name "Stones River group," upon concluding that the Central, Pierce, Ridley, and Glade limestones were equivalent to the Birdseye limestone of New York rather than to the New York Trenton, as Safford and Killebrew had believed. So close did Winchell and Ulrich regard the correlation of the Birdseye and Stones River beds that they then considered it likely that the name "Stones River" would eventually replace the nongeographic name "Birdseye." At that time Winchell and Ulrich considered the Birdseye older than the Black River, following the usage of Black River inaugurated by James Hall[8] in 1847.

In 1900 Safford and Killebrew[9] accepted the revived name "Stones River" and proposed formal geographic names for the Central and Glade limestones—Murfreesboro and Lebanon, respectively. The Carters Creek limestone was taken out of the Stones River group, the name shortened to "Carter," and the formation referred to the overlying Nashville group. In a table of formations[10] the Stones River group was classed with the Chazy and placed in a position to indicate equivalency of the group to the Lenoir limestone. Apparently this is the first record of the now time-honored correlation of the Chazy and Stones River.

Three years later, C. W. Hayes and Ulrich[11] put the Carter limestone back in the Stones River group and inaugurated the spelling "Carters."

Ulrich and Schuchert [12] expressed the opinion that most of the Stones River was pre-Birdseye. They considered the lower Stones River ". . . . about equivalent in time to the Chazy and Levis." Also, they maintained that the "extreme top of the Stones River is equivalent to the Lowville limestone."

Further changes in Ulrich's interpretation of the Stones River group appeared in 1908.[13] Cushing, in receipt of two letters from Ulrich—one dated November 19, 1907, and the other March 25, 1908—stated Ulrich's views as follows:

[5] *The Geology of Minnesota*, Vol. III (Part II, 1897), p. xc.

[6] *The Elementary Geology of Tennessee* (Nashville, 1876), pp. 130–33.

[7] P. xc of ftn. 5 (1897).

[8] "Description of the Organic Remains of the Lower Division of the New York System," *Paleon. N.Y.*, Vol. I (1847), pp. 14, 37, 46.

[9] *The Elements of the Geology of Tennessee* (Nashville, 1900), pp. 125–26.

[10] Safford and Killebrew, pp. 104–5 of ftn. 9 (1900).

[11] "Description of the Columbia Quadrangle, Tennessee," *U.S. Geol. Surv. Folio 95* (1903), correlation chart.

[12] "Paleozoic Seas and Barriers in Eastern North America," *N.Y. State Mus. Bull. 52* (1902), p. 641.

[13] H. P. Cushing, "Lower Portion of the Paleozoic Section of Northeastern New York," *Bull. Geol. Soc. Amer.*, Vol. XIX (1908), pp. 155–76.

> Ulrich no longer regards the Lowville as of upper Stones River age, but puts it above that formation. . . . . It should be stated also that the lower and middle divisions of the Stones River formation are regarded as essentially equivalent to lower and middle Chazy in time, but that the comparatively small Chazy basin was at this time entirely separated from the much larger sea to the south and west in which the Stones River deposits were laid down. "Finally the upper Stones River or Pamelia is only partially represented in the Chazy section by dove reef limestone at the base of the upper Chazy." [Last sentence quoted by Cushing from Ulrich's letter of March 25, 1908.]

Ulrich[14] reiterated his conclusion that the Pamelia, Crown Point, and Day Point limestones of New York were equivalent in time to the upper, middle, and lower divisions of the Stones River group of central Tennessee and that the Stones River group is therefore entirely Chazyan. He continued to regard the Carters limestone as the topmost formation of the Stones River group and made the Chazyan an Ordovician series composed of the Stones River and overlying Blount groups. Ulrich's Stones River group of eastern Tennessee was made up of the Mosheim (lower Stones River), Lenoir limestone (middle Stones River) and the upper, unnamed division, which was said to be developed along the western margin of the Appalachian Valley. By classifying the Mosheim and Lenoir limestones with the Stones River, Ulrich inaugurated the usage of the name "Stones River" for a widespread stratigraphic division of the Ordovician system. Thereupon the name also became a time term for the interval during which the Mosheim, Murfreesboro, Pierce, Ridley, Lenoir, Lebanon, and Carters formations were deposited.

R. S. Bassler[15] continued to regard the Mosheim as the basal formation of the Stones River. He took the Carters limestone out of the Stones River and made it the basal division of the Black River group. This change was adhered to by J. J. Galloway,[16] who also considered the Stones River to be Chazyan.

Butts[17] interpreted the Mosheim as older than the Stones River group of central Tennessee, in which he included only the Murfreesboro, Pierce, Ridley, and Lebanon. The most significant statement made by Butts about the Stones River concerns the supposed equivalency of the Lenoir and Ridley. He said:

> They [*Maclurea magna*] also serve to correlate the Lenoir unmistakably with the middle part of the Chazy limestone of the type locality in northeastern New York. *Maclurea magna* occurs rarely in the Ridley limestone of the Stones River group in middle Tennessee, and by this means the Lenoir is correlated with the Ridley.[18]

Bassler[19] had previously correlated the Lenoir with the Pierce, but as late as 1911 Ulrich[20] refrained from correlating the Lenoir with any single formation in the Central Basin. He pointed out that the Lenoir and the Stones River beds were deposited in separate basins with characteristic faunas which were not suitable for precise correlation.

[14] Revision of the Paleozoic Systems," *Bull. Geol. Soc. Amer.*, Vol. XXII (1911), p. 640, Pl. 27.

[15] "Bibliographic Index of American Ordovician and Silurian Fossils," *U.S. Nat. Mus. Bull. 92* (Vol. II, 1915), Pl. 1.

[16] "Geology and Natural Resources of Rutherford County, Tennessee," *Tenn. Geol. Surv. Bull. 22* (1919), pp. 30–31, 32–45.

[17] "The Paleozoic Rocks" in "Geology of Alabama," *Ala. Geol. Surv. Spec. Rept. 14* (1926), opp. p. 80.

[18] *Ibid.*, p. 105.

[19] Pl. 1 of ftn. 15 (1915).

[20] P. 539 of ftn. 14 (1911).

In 1928 Ulrich[21] and Butts concluded from a study of the section exposed along Yellow Branch, Lee County, Virginia, that the Mosheim was younger than the Murfreesboro. With this change, Ulrich's[22] revised Chazy classification was as follows:

Chazyan series
- Little Oak limestone (Alabama)

Blount group
- Ottosee shale and limestone
- Tellico sandstone
- Athens shale
- Whitesburg limestone
- Holston limestone

Stones River group
- Lebanon limestone
- Ridley-Lenoir limestones
- Mosheim limestone
- Murfreesboro limestone

This is the classification used by most geologists during the last fifteen years.

In 1939 Ulrich[23] abandoned his previous interpretation of the Chazyan and Stones River and introduced a new classification, shown in Figure 1. The disposition of the type Stones River formations is noteworthy. The Lebanon was placed in the Black River; and the Pierce-Ridley, long considered older than any part of Ulrich's Blount group, was placed above the Ottosee. Neither Blount nor Stones River appears in the classification. Two new names—Strasburg and Speers Ferry—were introduced, though without description or definition. This postulated succession negates Ulrich's prior contention regarding the existence of two Chazyan groups—the Stones River and overlying Blount. He continued to classify the Ridley and Pierce with the Chazy and maintained that the Murfreesboro is older than the Mosheim and Lenoir of the Appalachian Valley.

Although Ulrich, Bassler,[24] Butts,[25] and many others have held that the Stones River is Chazy, P. E. Raymond[26] has maintained since 1905 that the Stones River is post-Chazy. He stated:

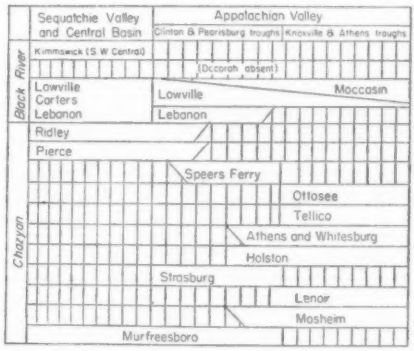

FIG. 1.—Correlation of Middle Ordovician formations in Tennessee (Ulrich, 1939).

Comparing the large percentage of forms common to the Stones River and to the Black River and Trenton with the low percentage—less than 5 per cent—of forms common to the Chazy and Trenton, it becomes evident that the Stones River and Trenton are faunally much more closely connected than are the Chazy and Trenton. These close relationships of the fauna of the Stones River to that of the Trenton, coupled with the stratigraphy, suggest

[21] "Ordovician Trilobites of the Family Telephidae and Concerned Stratigraphic Correlations." *Proc. U.S. Nat. Mus.*, Vol. LXXVI, Art. 21 (1929), p. 78.

[22] *Ibid.*, p. 73.

[23] "The Murfreesboro Limestone in Missouri and Arkansas and Some Related Facts and Probabilities," *Kan. Geol. Soc. 15th Ann. Field Conf. Guidebook* (1939), p. 106.

[24] "Stratigraphy of the Central Basin of Tennessee," *Tenn. Div. Geol. Bull. 38* (1932), pp. 48–58.

[25] Butts *et al.*, "The Southern Appalachian Region," *Guidebook 3, 16th Internat. Geol. Cong.* (1932), pp. 13–14; Butts, "Geologic Map of the Appalachian Valley of Virginia with Explanatory Text," *Va. Geol. Surv. Bull. 42* (1933), pp. 13–14, and "Geology of the Appalachian Valley in Virginia," *Va. Geol. Surv. Bull. 52* (Part I, 1940), pp. 119–78.

[26] "The Fauna of the Chazy Limestone," *Amer. Jour. Sci.*, Vol. XX (4th ser., 1905), pp. 353–82.

that the whole Stones River is younger than the Chazy.[27]

Schuchert[28] also believed that the Stones River is post-Chazy and stated:

> The writer made a list of 89 Stones River species, other than bryozoans. . . . . Of these 89 species, 7 are said to begin in the Chazy . . . . , and about 57 occur either in the Lebanon (a formation accepted by nearly every stratigrapher as of Black River age), or in other accepted Black River strata.

USAGE OF "STONES RIVER" IN THE APPALACHIAN REGION

The term "Stones River" has long been used in the Appalachian region for various successions of strata, generally including a fossiliferous zone characterized by *Maclurites magnus* Lesueur. In Pennsylvania, G. W. Stose[29] and Ulrich assigned 1,050 feet of beds in one section and at Martinsburg, West Virginia, 675 feet to the Stones River formation. A fauna collected from the middle, *Maclurites*-bearing division of his Stones River formation was interpreted by him to represent the middle Chazy of New York. He said:

> The fossils and the lithologic character of the formation are so nearly the same as those of the Stones River limestone of middle Tennessee that they are regarded as identical. Furthermore, the formation has been recognized by Ulrich and others at intervals along the Appalachian Valley southward into Tennessee. The name Stones River is therefore applied to this series of pure, fine-grained limestones.[30]

In Maryland, R. S. Bassler[31] assigned 1,100 feet in a generalized section to the Stones River. The middle fossiliferous division, as in Pennsylvania, contains *Maclurites magnus* and is about 200 feet thick. This part of the so-called Stones River is correlated by Bassler with the Crown Point limestone (middle Chazy) of New York.

In northern Virginia and particularly in the vicinity of Strasburg, Shenandoah County, Bassler[32] recognized 900 feet of beds in the Stones River; but, as Butts[33] has shown, most of this thickness is Beekmantown dolomite and limestone.

G. P. Grimsley[34] recognized a threefold division of the Stones River in the eastern Panhandle of West Virginia and gave the names "Lower, Middle, and Upper Stones River Limestone" to the three divisions previously recognized, but not named, by Stose in Pennsylvania.

D. B. Reger[35] has used "Stones River (Chickamauga) limestone" for the entire succession of limestones above the Beekmantown and below the Moccasin limestone (Fig. 2). According to Reger, the Stones River limestone in Mercer and Monroe counties contains "no vestige" of the lower Stones River (Murfreesboro).

In Virginia, Butts[36] has used the name "Stones River" for a succession ranging in thickness from a few feet to more than 1,300 feet, which directly overlies the Beekmantown and extends up to the top of the supposed Lenoir. The lowest division, occurring mainly south of New River and northwest of the Saltville

[27] *Ibid.*, p. 368.

[28] *Stratigraphy of the Eastern and Central United States* (New York: John Wiley & Sons, 1943), pp. 474–75.

[29] "Description of the Mercersburg-Chambersburg District," *U.S. Geol. Surv. Folio 170* (1909), pp. 7–8.

[30] *Ibid.*, p. 6.

[31] "Cambrian and Ordovician," *Md. Geol. Surv. Rept.* (1919), p. 120.

[32] "The Cement Resources of Virginia West of the Blue Ridge," *Va. Geol. Surv. Bull. II-A* (1909), p. 78.

[33] Pp. 197–98 of ftn. 25 (1940).

[34] "Jefferson, Berkeley, and Morgan Counties," *W.Va. Geol. Surv. Rept.* (1916), pp. 265–71.

[35] "Mercer, Monroe, and Summers Counties," *W.Va. Geol. Surv. Rept., 1926*, pp. 620–30.

[36] Pp. 119–47 of ftn. 25 (1940).

fault, was referred to as the "Blackford facies of the Murfreesboro formation." It contains characteristically reddish dolomitic siltstones with included pebbles of chert and dolomite, light-gray mealy shales, and fossiliferous cherty layers. In some places, particularly at St. Clair Station, Tazewell County, Virginia—less than 4 miles from Mercer County, West Virginia, wherein Reger[37] believed the Murfreesboro to be absent—Butts recognizes a locally developed limestone facies of the Murfreesboro, with a thickness of more than 1,000 feet. Beds which he believed to represent the Lenoir and Mosheim were also classified[38] with the Stones River, and the Lenoir was correlated with the Ridley of central Tennessee. Butts correlated the Murfreesboro with the basal member of the Chazy of Champlain Valley in New York and the Lenoir with the Crown Point limestone.

In summary, "Stones River" as a formation and group name has been used widely in the Appalachian region, but the various usages are inconsistent. In spite of the many similarly emphatic correlations of beds in the Appalachian region with the Stones River group, very little faunal evidence has been presented to substantiate the correlations. Until recently,[39] Murfreesboro, wherever used

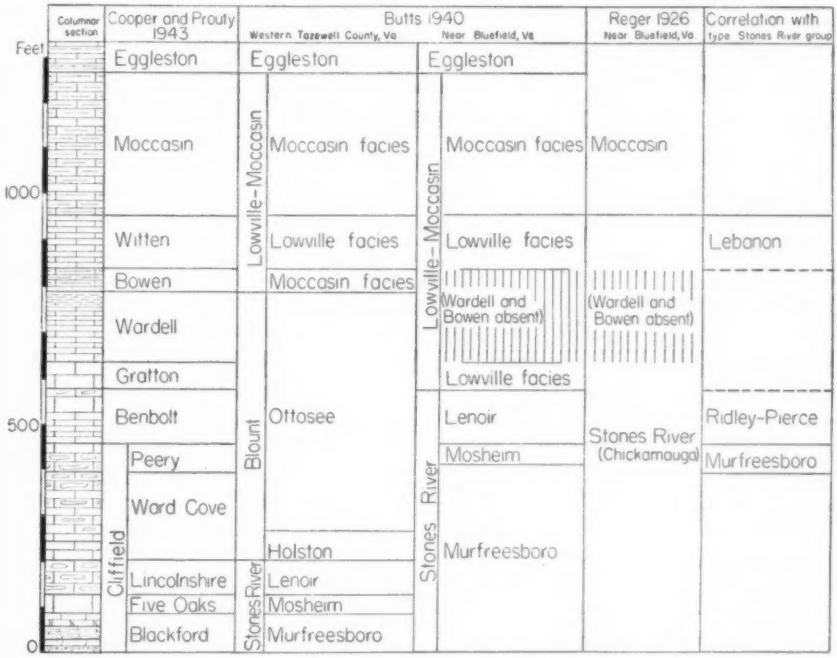

FIG. 2.—Position of Stones River equivalents in belts northwest of Clinch Mountain in southwestern Virginia.

[37] P. 622 of ftn. 35 (1926).

[38] Butts, p. 147 of ftn. 25 (1940).

[39] B. N. Cooper and C. E. Prouty, "Stratigraphy of the Lower Middle Ordovician of Tazewell County, Virginia," *Bull. Geol. Soc. Amer.*, Vol. LIV (1943), pp. 819–86; B. N. Cooper, "Geology and Mineral

in the Appalachian region, has been generally regarded as the oldest post-Beekmantown formation. The Carters is no longer considered part of the Stones River. Various classifications of the Stones River formations are shown in Table 1.

STRATIGRAPHIC POSITION OF THE MURFREESBORO IN CENTRAL TENNESSEE

In the original description of the formation[40] the Murfreesboro was described as the oldest limestone in the Central Basin and as consisting of the lowest 70 feet of exposed beds. Ulrich[41] regarded the type Murfreesboro as probably resting on the Beekmantown or Knox. The true position of the type Murfreesboro did not become generally known until 1940, when K. E. Born[42] published a detailed description of the cuttings obtained from the Basin Oil and Gas Company's Henry Harrell No. 1 well, which was drilled in the circular outcrop area of the Murfreesboro at the type locality, Rutherford County, Tennessee. He identified 380 feet of limestone above the Knox and below the type Murfreesboro. Thus Ulrich's supposition that the Murfreesboro of the type locality is probably just above the Beekmantown is erroneous. The true position of the type Murfreesboro with respect to the subsurface Knox dolomite is of special interest because of the remarkably similar position of the Murfreesboro fauna in the Ordovician succession of southwestern Virginia.

Resources of the Burkes Garden Quadrangle, Virginia," *Va. Geol. Surv. Bull. 60* (1944), p. 71.

[40] Safford and Killebrew, p. 125 of ftn. 9 (1900).

[41] Pl. 27 of ftn. 14 (1911).

[42] "Lower Ordovician Sand Zones ('St. Peter') in Middle Tennessee," *Bull. Amer. Assoc. Pet. Geol.*, Vol. XXIV (1940), pp. 1641–62.

IN THE APPALACHIAN REGION

Midway between Wittens Mills and Five Oaks, Tazewell County, Virginia, the Peery member of the Cliffield formation[43]—400–475 feet above the Beekmantown dolomite—contains an assemblage of fossils found in the type Murfreesboro (Fig. 2). Among the species identified from the cherty beds of the Peery member are *Lophospira procera* Ulrich, *L. perangulata* (Hall), *Helicotoma* aff. *H. tennesseensis* Ulrich and Scofield, *H. declivis* Ulrich, *Trochonema bellulum* Ulrich, *Ophiletina sublaxa depressa* Ulrich and Scofield, *Plectoceras bondi* (Safford), *Tetradium syringoporoides* Ulrich, *Leperditia fabulites* (Conrad), *L. pinguis* Butts, *Phragmolites* sp., and peculiar forms resembling closely *Polylopia billingsi* (Safford). In the same section Butts recognizes the Murfreesboro, Mosheim, and Lenoir (Fig. 2), but the top of his Lenoir is at least 200 feet stratigraphically below beds containing the Murfreesboro fossils mentioned above. Also in this section, the beds comprising the upper part of the Peery limestone member are a typical calcilutyte,[44] like the Mosheim. At two other localities in Tazewell County—one at Pounding Mill[45] and the other at St. Clair—this calcilutyte has been identified by Butts[46] as Mosheim. That this zone of Mosheim lithology is several hundred feet above the limestone identified by Butts as Mosheim in the section near Wittens Mills has been shown previously.[47] It can also be positively affirmed that the cherty limestones with Murfreesboro

[43] Cooper and Prouty, pp. 866–68 of ftn. 39 (1943).

[44] *Ibid.*, p. 826.

[45] Butts *et al.*, p. 65 of ftn. 25 (1932).

[46] Pp. 123–24 of ftn. 25 (1940).

[47] Cooper and Prouty, pp. 852–53 of ftn. 39 (1943).

TABLE 1

CLASSIFICATIONS OF THE STONES RIVER GROUP

Safford, 1851	Safford, 1869	Safford and Killebrew, 1876	Ulrich and Winchell, 1897
Stones River group: Upper Lebanon limestone	Trenton (Lebanon) formation: Carter's Creek limestone	(Eastern Tenn.) Nashville group	Stones River group: Carter's Creek limestone
Lower Lebanon limestone	Glade limestone	Lebanon group: 380 feet of marble in the Knoxville area	Glade limestone
	Ridley limestone		Ridley limestone
	Pierce limestone		Pierce limestone
Stones River beds	Central limestone	Lenoir limestone	Central limestone

Safford and Killebrew, 1900	Ulrich and Hayes, 1903	Ulrich, 1911 (Generalized Section)	Ulrich, 1911 (Eastern Tenn.)
Carter limestone	Stones River group: Carters limestone	Stones River group: Carters limestone	Lowville limestone
Stones River group: Lebanon limestone	Lebanon limestone	Lebanon limestone	Stones River group: Upper Stones River
Ridley limestone	Ridley limestone	Ridley limestone	
Pierce limestone	Pierce limestone	Pierce limestone	Lenoir limestone
Murfreesboro limestone	Murfreesboro limestone	Murfreesboro limestone	
		Mosheim limestone	Mosheim limestone

Bassler, 1915	Galloway, 1919	Butts, 1926	Ulrich, 1929
Carters limestone	Carters limestone	Carters limestone	Lowville limestone
Stones River group: Lebanon limestone	Stones River group: Lebanon limestone	Stones River group: Lebanon limestone	Stones River group: Lebanon limestone
Ridley limestone	Ridley limestone	Ridley limestone	Ridley limestone
Pierce limestone	Pierce limestone	Pierce limestone	Pierce limestone
Murfreesboro limestone	Murfreesboro limestone	Murfreesboro limestone	Mosheim limestone
Mosheim limestone		Mosheim limestone	Murfreesboro limestone

TABLE 1—*Continued*

Bassler, 1932		Butts, 1940 (Appalachian Valley)		Butts, 1940 (Central Tenn.)		Ulrich, 1939 (Central Tenn.)	
Carters limestone		Blount group		Carters limestone		Black River group	Lowville limestone
Stones River group	Lebanon limestone	Stones River group	Lenoir limestone	Stones River group	Lebanon limestone		Carters limestone
							Lebanon limestone
	Ridley limestone		Mosheim limestone		Ridley limestone	Chazyan	Ridley limestone
	Pierce limestone				Pierce limestone		Pierce limestone
					Hiatus?		*Major hiatus*
	Murfreesboro limestone		Murfreesboro limestone		Murfreesboro limestone		Murfreesboro limestone

fossils are at least 200 feet above the *Sowerbyites triseptatus–Dinorthis atavoides* zone, which, with few exceptions, constitutes Butts's Lenoir limestone in Tazewell, Russell, and Scott counties of southwestern Virginia.

Along Yellow Branch, Lee County, Virginia, Butts[48] reported Murfreesboro fossils from a thin zone 175 feet above the Beekmantown. This zone is part of a 100-foot succession between two almost equally thick zones of calcilutyte. The upper calcilutyte was identified by Butts and Ulrich as the Mosheim, and this identification is the basis for their claim that the Mosheim is younger than the Murfreesboro. Since two similar zones of calcilutyte have been confused with the Mosheim in some other localities in southwestern Virginia, there is justification for doubting the identity of the Mosheim as determined by Butts and Ulrich in the Yellow Branch section. Butts's Mosheim of the Yellow Branch section occupies the same stratigraphic position as the calcilutyte at the top of the Peery limestone member, being above beds containing the Murfreesboro fauna and directly below argillaceous layers containing *Öpikina*, *Mimella*, and bryozoans such as occur in the Pierce and Ridley of central Tennessee. This calcilutyte near Wittens Mills is part of Butts's Ottosee. In view of the confusion regarding the identity and position of the Mosheim in southwestern Virginia, it can no longer be regarded as a certainty that the Mosheim is younger than the Murfreesboro.

One mile north of Ceres, Bland County, Virginia, the Peery limestone overlies a succession of black limestones and shales containing the *Nemagraptus gracilis* fauna characteristic of a formation identified by Butts as Athens (Fig. 3). The Murfreesboro, which is the equivalent of the Peery, is therefore younger than most of Ulrich's Blount formations. According to Ulrich, the Murfreesboro belongs at the base of the limestone section—hundreds of feet below the Athens!

It is difficult to establish precisely the position of the Murfreesboro with respect to the type Mosheim and Lenoir, since all three have not been positively identified in any section. The *Sowerbyites triseptatus–Dinorthis atavoides* zone—200-

[48] Pp. 120–21 of ftn. 25 (1940).

250 feet below the true Murfreesboro equivalent in the section near Wittens Mills—is Ulrich's "Strasburg (upper Lenoir)" and Butts's Lenoir in the also occurs in limestone intercalations in dolomitic chert-breccias of Butts's "Blackford facies of the Murfreesboro formation" along Moccasin Creek, south

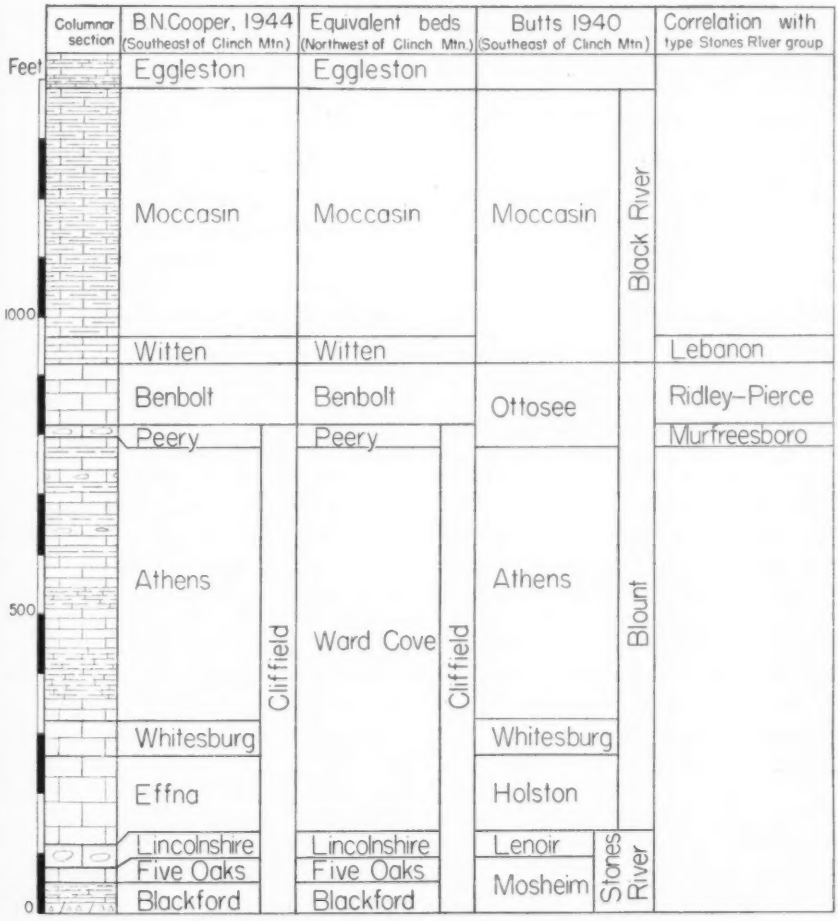

FIG. 3.—Position of Stones River equivalents in belts southeast of Clinch Mountain in southwestern Virginia.

Clinch Mountain belt of Virginia and Tennessee. Whether this zone is a part of the type Lenoir is unknown; probably the typical Lenoir is older. The lowest beds of the type Lenoir are crowded with *Rostricellula pristina* (Raymond), which of Tumbez, Russell County, Virginia. A normal development of the *Dinorthis atavoides–Sowerbyites triseptatus* zone begins about 175 feet higher in the Tumbez section. If the *Sowerbyites-Dinorthis* zone is part of the type Lenoir, most, if

not all, of the succession down to the Beekmantown in southwestern Virginia could be correlated with the Lenoir. The Murfreesboro equivalent is definitely placed considerably above the *Sowerbyites-Dinorthis* zone near Wittens Mills and in Bland County succeeds the Athens formation of Butts, which is supposed to be considerably younger than the type Lenoir.

STRATIGRAPHIC POSITION OF THE PIERCE-RIDLEY IN THE APPALACHIANS

The Benbolt limestone, which succeeds the Peery limestone in southwestern Virginia, contains many fossils found in the Pierce and Ridley of the Central Basin. Perhaps the most apparent and convincing similarity is the abundance of strophomenoid brachiopods in the three formations. *Öpikina*, particularly *Ö.* cf. *Ö. minnesotensis*, and *Strophomena* are prominent. *Camerella plicata* Schuchert and Cooper and similar forms of *Campylorthis*, *Glyptorthis*, *Paurorthis*, and *Hesperorthis* also occur in the three formations. "*Protorhyncha ridleyana*," recently redefined and renamed *Ancistrorhyncha costata* Ulrich and Cooper and usually regarded as a Ridley guide fossil, was collected by G. A. Cooper from beds in Rye Cove, Scott County, Virginia, which are referred by the writer to the Benbolt and classed by Butts with the Ottosee. The same, or closely allied, forms of the bryozoan genera *Anolotichia*, *Chasmatopora*, *Graptodictya*, *Nicholsonella*, and *Rhinidictya* occur in the Benbolt and in either the Pierce or Ridley. The position of the Benbolt, directly above the Murfreesboro equivalent in Tazewell County, Virginia, is corroborative evidence for the identification of the Benbolt as the equivalent of the Pierce and Ridley. *Fascifera subcarinata* and *Stromatocerium rugosum*, which occur in the Ridley, are also found in the lower part of the Wardell formation (upper Ottosee) in southwestern Virginia and along U.S. Route 25E, about $1\frac{1}{2}$ miles southeast of Evans Ferry, Grainger County, Tennessee. Thus the Benbolt probably does not represent all the Ridley.

STRATIGRAPHIC POSITION OF THE LEBANON

The equivalent of the Lebanon limestone of central Tennessee is the Witten limestone,[49] which corresponds to Butts's Lowville along Clinch Mountain in southwestern Virginia and Tennessee. Not only is the general lithology of the Lebanon and Witten much the same, but many fossils are common to the two. Among these are *Cryptophragmus antiquatus* Raymond, *Pionodema minuscula* Willard, similar forms of *Sowerbyella*, *Rostricellula*, *Doleroides*, and *Zygospira*,[50] and abundance of bryozoans of the genera *Escharopora*, *Orbignyella*, *Rhinidictya*, and *Nicholsonella*. Also similar, if not identical, types of fucoids and *Camarocladia*-type sponges are found in both the Lebanon and Witten.

In southwestern Virginia the Benbolt and Witten are separated by 200–400 feet of beds, comprising the Gratton, Wardell, and Bowen formations (Fig. 2). Possibly part of this succession may be represented in the upper Ridley and lower Lebanon of central Tennessee; but, judging from the greater thickness and the greater diversity of faunas, the Benbolt-Witten succession in southwestern

[49] Cooper and Prouty, pp. 877–79 of ftn. 39 (1943).

[50] Information on the occurrence of *Cryptophragmus antiquatus*, *Pionodema minuscula*, and other fossils in the Lebanon was furnished by G. Arthur Cooper, who, in company with Josiah Bridge, collected them from the middle part of the Lebanon limestone in central Tennessee.

Virginia would seem to represent a more nearly complete record of deposition than the Pierce, Ridley, and Lebanon of central Tennessee. A hiatus between the Ridley and Lebanon, therefore, seems probable.

AGE OF THE STONES RIVER BEDS

Ulrich's conception of the Chazyan series in the Appalachian region, embracing the Stones River and Blount groups, is untenable. The Murfreesboro is considerably younger than the Mosheim and Lenoir, and it is above the *Nemagraptus gracilis* zone that occurs in the Athens of Virginia and the Normanskill of New York. Most geologists have considered the typical Lenoir to be equivalent to the Crown Point limestone of the middle Chazy of New York. P. E. Raymond[51] concurs in this correlation and assigns the Normanskill-Athens graptolite zone to the upper Chazy. Although Ulrich places the Ottosee in the Chazy, no faunal basis for the correlation was ever given, and none is known to exist. Although it is very doubtful whether beds as young as the Athens are actual correlatives of the upper division of the Chazy, this correlation could be accepted without regarding any of the Stones River as Chazyan. Now that the Ridley and Lenoir are known not to be equivalent, the main "basis" for linking the Stones River with the Chazy is removed.

The fossils in the Stones River formations and in their Appalachian equivalents strongly suggest a post-Chazyan age. The main elements in the brachiopod faunas, particularly species of *Öpikina*, *Campylorthis*, *Fascifera*, *Hesperorthis*, *Doleroides*, *Zygospira*, and *Strophomena*, are characteristic of formations generally conceded to be Black River in the Mississippi Valley region. Other Stones River fossils considered to be post-Chazy include *Stromatocerium rugosum*, *Foerstephyllum halli*, *Fletcheria*, and *Lambeophyllum*—all of which occur in the Benbolt-Witten and Pierce-Lebanon successions. Even *Tetradium syringoporoides*, long regarded as an infallible index of the Chazy, is reported to be common in the lower Black River of New York and Ontario.[52] *Maclurites magnus*, which is widely considered as an index of the Chazy, has been reported from the Murfreesboro, Pierce, Ridley, and Lebanon of central Tennessee[53] and from beds identified by Butts as Ottosee.[54] Either this species has long range or it has been loosely identified. A few other species, notably *Leperditia fabulites* and *Lophospira perangulata*, which occur in the Stones River, are reported from the Chazy; but they have been cited also from the Pamelia and Lowville. As previously indicated by Raymond and Schuchert, the dominance in the Stones River of forms essentially foreign to the Chazy and characteristic of post-Chazy formations favors a post-Chazy age for the entire Stones River group of central Tennessee.

Cryptophragmus antiquatus, which is common in the Pamelia and lower Lowville of New York, occurs in association with other Black River fossils in the Witten and Lebanon limestones. If the Lebanon and equivalent beds in the Appalachian region are Black River and if

[51] "Correlation of the Ordovician Strata of the Baltic Basin with Those of Eastern North America," *Bull. Harvard Mus. Comp. Zoöl.*, Vol. LVI, No. 3 (1916), pp. 236–37.

[52] V. J. Okulitch, "Some Black River Corals," *Trans. Royal Soc. Canada*, Vol. XXXII (3d ser., 1938), pp. 93–94.

[53] Bassler "Stratigraphy of the Central Basin of Tennessee," *Tenn. Div. Geol. Bull. 38* (1932), pp. 51–58.

[54] Cooper and Prouty, pp. 868–71 of ftn. 39 (1943).

the entire Stones River group is post-Chazy, a new lower Middle Ordovician group should be recognized. Although the existence of such a division is strongly indicated by what is already known of the fossils and stratigraphy of the lower Middle Ordovician in southwestern Virginia, the precise lower limit is still in question and must remain so until the upper limit of the Chazy in the Appalachian region is accurately determined. Although New York is certain to remain the "hub" of Ordovician correlations, the succession in that area must not be adjudged necessarily "ideal" for the whole United States.

Since the stratigraphic relations of the type Murfreesboro to unexposed older beds is obscure and since an integral, if not the most characteristic part, of the Stones River group—the Lebanon limestone—seems unmistakably linked with the Black River, there is no reason for continued usage of the name "Stones River group." If the Ridley, Pierce, and Murfreesboro are to be recognized as a part of a new post-Chazyan, pre-Black River group, the name "Stones River" should not be revived. The new group should be named and defined from an Appalachian section, where relations with subjacent and overlying formations are fully apparent.

SUMMARY OF CONCLUSIONS

1. The Murfreesboro, long considered the oldest post-Beekmantown formation in the Appalachian region, is considerably younger than the Mosheim and Lenoir. In Bland County, Virginia, the Peery limestone, which is the faunal equivalent of the Murfreesboro, overlies a formation of black limestones and shales containing the *Nemagraptus gracilis* fauna which is characteristic of Butts's Athens in southwestern Virginia.

2. The Benbolt limestone, which immediately succeeds the Peery limestone in Bland and Tazewell counties, Virginia, correlates with the Pierce and Ridley of central Tennessee.

3. The Witten limestone of southwestern Virginia, constituting a part of what Butts has called "Lowville" limestone, is faunally linked with the Lebanon of the Central Basin.

4. Since the Benbolt and Witten are separated by three formations, two of them with rather distinctive faunas, the succession of southwestern Virginia would seem to represent a fuller record of deposition than the Pierce-Lebanon of the Central Basin. An unconformity probably exists between the Lebanon and Ridley in the latter area.

5. Confusion regarding the relative ages of the Murfreesboro, Mosheim, and Lenoir resulted from failure to recognize the occurrence of more than one zone of "characteristic" Mosheim lithology. Two such zones have been called "Mosheim" in southwestern Virginia. It seems to be clearly demonstrated that the bed identified as "Mosheim" along Yellow Branch, Lee County, is far above the position of the type Mosheim.

6. Since the Murfreesboro is younger than the Lenoir, the Ridley and Lenoir cannot be equivalent.

7. Ulrich's concept of a "Chazyan series," embracing the Stones River and overlying Blount groups, is invalid. The Mosheim and Lenoir are older than Stones River, and the Stones River formations are younger than the greater part of Ulrich's Blount group.

8. Fossils purported to link the Stones River with the New York Chazy are either forms known to have long range or ones which probably have been loosely identified. They have little index value.

9. The predominance of post-Chazyan

forms in the Stones River beds and in equivalent formations in the Appalachian region would strongly suggest that the entire Stones River group is younger than the Chazy. The uppermost Stones River formation, the Lebanon, seems clearly to represent part of the Black River.

10. The available evidence seems to indicate the probable existence of a post-Chazy, pre–Black River group in the Appalachian region—a group which includes beds at least as old as the Peery limestone and possibly as young as the Bowen or the Wardell formations. The Murfreesboro, Pierce, and Ridley would represent a substantial part, though probably not all, of this group. The precise limits of the division have not yet been determined, for lack of sufficient information about the fossils and about the position of the upper Chazyan boundary in the Appalachian section.

11. "Stones River" as a general time-rock term should no longer be used, nor should the name be revived if a post-Chazyan, pre–Black River group is recognized.

ORIGIN OF GRANITE DOMES IN THE SOUTHEASTERN PIEDMONT

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ABSTRACT

The forms of the granite domes of the Southeast, which heretofore have been attributed to exfoliation, are regarded as the product of granular disintegration brought about by chemical weathering.

INTRODUCTION

There are many dome-shaped exposures of granite throughout the Southeastern Piedmont. The largest, the well-known Stone Mountain of De Kalb County, Georgia, is well known; but its many smaller counterparts throughout the Piedmont of Georgia and the Carolinas are very little known. The domes vary greatly in size from the massive Stone Mountain itself, which is about a mile and a half long and some 650 feet higher than the surrounding peneplain, down to small inconspicuous bosses, little larger than residual boulders. They are all alike, however, in having smooth, spheroidal surfaces without sharp protuberances or re-entrants. Most are composed of granite, but some few are gneissic. None show visible joint systems.

Heretofore they have been explained as the product of exfoliation; but the writer, after several years of field work throughout the region in which they occur, has come to believe that they have resulted largely from the action of other processes. The validity of the exfoliation theory of origin for granite domes in certain other climatic regions is not questioned, but those which have developed from the granites and gneisses of the southeast do not show enough evidence of exfoliation to justify the assumption that it has played a dominant or even an important part in their formation. It is true that some of the southeastern domes show a little exfoliation; but in most instances the surface is smooth and unbroken, and evidence of exfoliation is rarely seen (Fig. 1).

The exfoliation theory of dome development stands upon the essentially sound reasoning that an unjointed mass of homogeneous rock will be attacked most readily on those parts of its surface which have the smallest radii of curvature and that the continuation of such selective attack will reduce the mass to a spheroidal form. But this principle applies equally well to the denudation caused by any other weathering agent which is nondirectional in its attack. In the southeastern states the climate is warm and humid, and the remarkable development of chemical weathering has long been a matter of comment. Hydration, oxidation, and carbonation are nondirectional in their attack and should be adequate to reduce an irregular-shaped mass to spheroidal form provided the rock resistance is uniform. Since the granites and gneisses from which the domes have been sculptured are quite homogeneous and without joints, there is little reason to believe that their resistance to chemical weathering should be differential.

While chemical weathering has usual-

ly been most pronounced in areas underlain by jointed or schistose rocks, there is good evidence that it has also had a significant effect upon the unjointed masses of granite and gneiss from which these domes have been shaped. Quarrymen working on them usually find it necessary to remove 6 inches to several feet of sap before they encounter unaltered rock. These are not impressive thicknesses, but kaolinization characteristically takes place on the intergranular surfaces of the feldspars, and the rock at the surface of the exposure tends to break up into a gruss which is washed off the steep bare slopes as quickly as it is formed. Many observers have noted this intergranular alteration. L. E. Smith[1] has made petrographic studies showing it in the surficial phases of the unjointed granite masses of the South Carolina Piedmont, and the writer has made similar studies in North Carolina.

FIG. 1.—Stone Mountain in Wilkes County, North Carolina.

EVIDENCE FOR EXFOLIATION

It is evident that exfoliation and granular disintegration are coexistent as sculpturing agents on the domes, and their relative importance should be determined by a comparison of the evidence for each. Considering first the case for exfoliation, if a dome were dominantly the result of its action, one would expect to find that fact manifest in two ways. (1) Since few exfoliation spalls detach themselves from the parent-mass in the form of complete lenses, one would expect to find the surface of the dome covered by truncated remnants of spalls which had partially fallen away. There would probably be overlapping of such remnants, and the surface of the dome would have a somewhat imbricate appearance. Half Dome in Yosemite (Fig. 2) offers an excellent example of such a surface. (2) If the surface of the dome had been produced by exfoliation, at its base one would expect to find a talus slope composed of fallen and broken spalls.

In the case of the domes of the Southeast neither of these criteria is satisfied. As stated above, there is very little evidence of exfoliation on the dome surfaces. Looking at the face of Stone Mountain in Wilkes County, North Carolina, the observer can see the broken edges of no more than two or three spalls from any one viewpoint. And at the base of the mountain only a few remnants of fallen spalls can be found. This observation applies equally well to the attendant bosses and minor domes which appear near by. On Stone Mountain in De Kalb County, Georgia, there is somewhat better evidence of exfoliation; and it seems to have been a slightly more important factor in the denudation of that mass. Even there, however, the broken edges of spalls are so rounded by normal weathering (granular disintegration) that they are not conspicuous. Other less well-known domes throughout the region

[1] "Weather Pits in Granite of the Southern Piedmont," *Jour. Geomorph.*, Vol. II (1941), p. 125.

show the effects of exfoliation in varying intensity, but most of them are as little affected by it as is Stone Mountain in Wilkes County, North Carolina. Some are affected even less.

It is true that there seems to be a latent tendency toward hypogene exfoliation in all the domes, but there does not appear to be much evidence that it has ever been activated by wholly natural agencies. In every place where it can be recognized, it has been initiated artificially by the rapid removal of overburden in the process of quarrying. Most of the natural spalls are quite thin and seem to be the result of other causes.

It is possible that the domes were produced by exfoliation under different climatic conditions in a previous geologic age. However, this possibility seems remote, for there is no evidence of an arid climate in this region later than Triassic time.

In brief, if a dome were sculptured by exfoliation, the evidence for the action of that agency should be spectacularly displayed as remnants of broken spalls both adhering to the dome surface and composing a talus slope at its base.

FIG. 2.—Half Dome in Yosemite Valley, California, after F. E. Matthes. Photograph by courtesy of the U.S. Geological Survey.

EVIDENCE FOR GRANULAR DISINTEGRATION

On the other hand, if a dome had developed through granular disintegration, the evidence to prove it should be somewhat obscure. Formation of gruss is a grain-by-grain process. As soon as a grain has been loosened from the parent-mass, it is washed off the steeply sloping surface either to become part of the bed load of the small drainage ways at the

base of the dome or to be incorporated in alluvial fans surrounding it. Because of the slow rate at which the individual grains are released from the parent-mass, there is small tendency for fans to develop; but in every location which is favorable to the detention of sediment there are deposits of the coarse debris produced by granular disintegration. Furthermore, every dome of the writer's acquaintance is surrounded for significant distances by the young soils which result from granular disintegration. The dominant series is the Louisburg, a light-colored sandy soil without definite profile development.

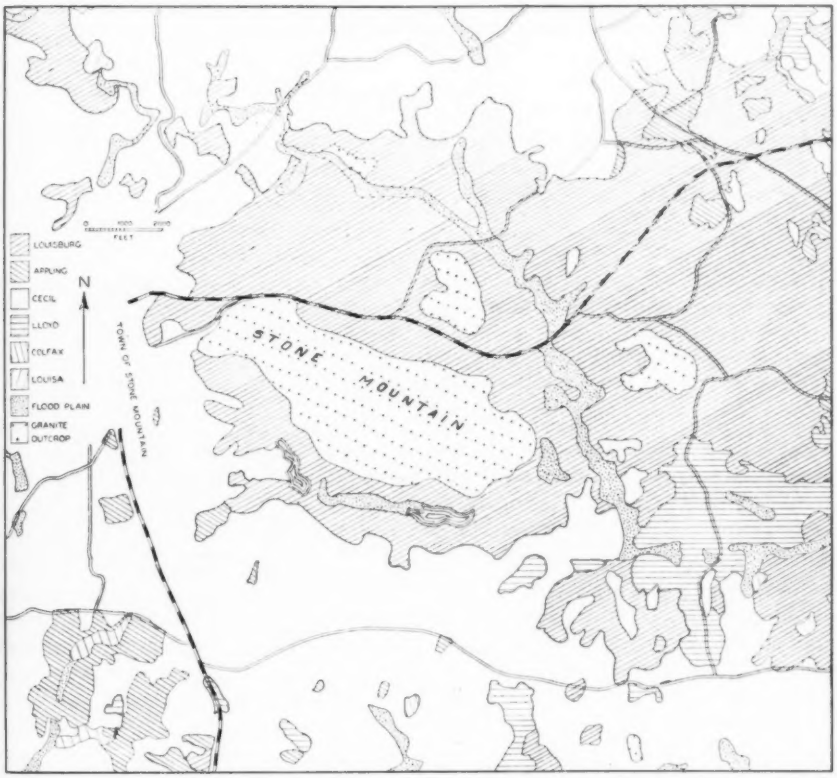

FIG. 3.—Map of area around Stone Mountain in De Kalb County, Georgia, showing distribution of soil series.

Field evidence[2] suggests that it develops largely upon unjointed granites and gneisses as the product of granular disintegration. The intimacy of its association with the domes is exemplified by the soils map shown in Figure 3. This has been reproduced from unit-area maps of the Soil Conservation Service, United States Department of Agriculture,[3] and

[2] W. A. White, "Determining Factors in the Coloration of Granite Soils in the Southeastern Piedmont," *Amer. Jour. Sci.*, Vol. CCXLII, No. 7 (1944), pp. 361–63.

[3] P. H. Montgomery, "Erosion and Related Land Use Conditions of the Lloyd Shoals Reservoir Watershed, Georgia," Phys. Surv. Div., Soil Cons.

shows the area surrounding Stone Mountain in De Kalb County, Georgia. It will be noted that the dome is surrounded by the young Louisburg series, although the region in general, away from the dome, is overlain by the more common Cecil—a mature series which, as the writer has shown elsewhere,[4] characteristically develops on closely jointed acid rocks. The area covered by Louisburg extends for some distance to the east of Stone Mountain, where it surrounds a number of smaller low domes or "flat rocks." Two of these are shown on the map, to the northeast of Stone Mountain.

On sheet 64 of the same series of maps from which Figure 3 was reproduced a similar distribution of Louisburg soil may be seen surrounding Pine Mountain, which is located about 1 mile east of Lithonia, Georgia. This is a smaller dome than Stone Mountain and has been reduced to a lower profile, but it also gives every manifestation of having been produced by granular disintegration rather than exfoliation. Many other small domes in all stages of reduction appear throughout the area east of Stone Mountain, Georgia. All of them are surrounded by Louisburg soil and show little evidence of natural exfoliation.

VALLEY-WALL FLAT ROCKS

In general, the upland surface of the Piedmont is a peneplain which is very poorly developed along its western edge but shows increasing development to the east. At its easternmost limit, near the edge of the Coastal Plain, it has been reduced to a very low relief, and monadnocks are rare or absent. There has been uplift and some dissection. The domes differ greatly in topographic age, as measured by their local relief. In the upper Piedmont, where peneplanation is not as well developed, they are largely mature with high profiles and steep sides. In the middle Piedmont they are in general older and of lower relief, many being almost flat. Near the Coastal Plain, where peneplanation was most highly developed, their former presence is represented by exposures along valley walls where dissection is taking place.

These last are characteristically wide, slightly inclined, and broadly curved exposures of unjointed granite with surfaces unbroken save for occasional weather pits. Uphill they pass under residual soils of the Louisburg or Durham series, which in some places are capped by a veneer of marine deposits. Typical examples are Flat Rock and Forty-Acre Rock near Kershaw, South Carolina.

From their topographic positions in the walls of young valleys and from the fact that they pass under residual soil, one draws the conclusion that the present exposures of these flat rocks were sculptured by a later erosion cycle than that which shaped the monadnock domes higher in the Piedmont. However, since these areas were highly peneplaned, it would seem plausible to believe that these valley-wall flat rocks merely represent new exposures of former domes which were reduced by peneplanation to the point where they could develop a residual soil. Geomorphically, such former domes would have been of the same generation as the present examples found in the upper Piedmont, although chronologically older; that is, because the peneplain developed first in the lower Piedmont and extended progressively westward, the former monadnock domes which developed on the lower edge of it

Serv., U.S. Dept. Agric., 1940. Parts of sheets 33, 34, 46, and 47.

[4] White, pp. 361–63 of ftn. 2 (1944).

had been reduced to the general level and covered by residual soil before the present dissection again exposed their flanks as valley-wall flat rocks. On the other hand, the present monadnock domes of the upper Piedmont developed later with the western migration of peneplanation, and they have not yet been reduced to the general level.

A continuation of this reasoning would suggest that many of the areas of Louisburg and Durham soil which are found in undissected parts of the Piedmont also indicate the location of former domes which have been reduced to flat-rock status and buried by residual soil.

In those instances where valley-wall flat rocks pass under remnants of Tuscaloosa or "Lafayette" deposits, it is, of course, possible that former low domes were planated by wave erosion. However, the absence of any wave-cut scarp at the edge of the Coastal Plain argues against this possibility and suggests that peneplanation had been extreme before the last submergence.

Where valley-wall flat rocks pass under marine terrace deposits, it is obvious that the localization of dissection has been the result of superimposition from consequent drainage. Therefore, the development of spheroidal surfaces on the granites exposed in the walls of such narrow valleys demonstrates that domes can develop independently of any control which might be exerted by internal structures of the granite itself, such as concentric strain lines, schlieren, or petrofabric orientation. In several such places, as at Forty-Acre Rock near Kershaw, South Carolina, spheroidal surfaces appear on both sides of the dissecting valley—a fact which argues further against the influence of internal structures, for the directions of curvature are reversed on opposite sides of the valley.

INFLUENCE OF INDURATED VENEERS

Another factor which is believed to have had some significance in giving the domes their smooth spheroidal surfaces may be found in the indurated veneers which characteristically appear on their exposed surfaces.[5] These veneers seem to result from the deposition and oxidation of iron compounds which have been carried upward in solution by capillary water. Deposition takes place near the surface when the water evaporates and serves to reconsolidate the partially disaggregated mineral grains of the altered sap rock. Such indurated veneers on the domes tend to slow weathering and topographic reduction both by restricting the entrance of air and surface water to the underlying rock and by increasing the resistance of the surface to mechanical disintegration. However, they are of most influence in the development of topographic form when they appear on flat-rock exposures in areas subject to dissection. There, by their tendency to reduce weathering, they increase the differential in rate of decomposition between the exposed rock and that underlying the surrounding soil.

When dissection takes place, this sharply marked distinction between altered and unaltered rock at the edge of the old exposure localizes the steep edge of a table rock of the type shown in Figure 4, in which a veneer can be seen preserving the surface of a former small flat rock. After prolonged exposure such a form will lose the sharp protective edges of its original veneer; but, as the edges are rounded, it will develop an extension of the veneer down its sides and resolve at length into a small dome. This new ex-

[5] White, "Geomorphic Effects of Indurated Veneers on Granites in the Southeastern States," *Jour. Geol.*, Vol. LII (1944), pp. 333–41.

tension to the original veneer will have similar protective qualities and will tend to preserve the sharp boundary between the dome and the surrounding soil. This process may possibly explain the sharp knick points which characteristically appear at the bases of domes, and (although the writer does not like to extend his conclusions to regions unfamiliar to him) it may have some connection with the genesis of the bornhardts of East Africa.

In conclusion, the writer would like to suggest that perhaps geologists have erroneously considered granite domes to be unusual land forms, produced only by very special conditions. It would probably be more catholic to regard them as the expected form wherever nonjointed homogeneous rocks are subjected to the attack of any nondirectional agency of denudation. The particular agency would be determined by the local climate, but the resulting land forms should be essentially similar.

FIG. 4.—Granite outcrop protected by indurated veneer, 2 miles southwest of Wendell in eastern Wake County, North Carolina.

REVIEWS

The Pleistocene Geology of Iowa: Special Report. By GEORGE E. KAY AND OTHERS. Part I: "The Pre-Illinoian Pleistocene Geology of Iowa," by GEORGE F. KAY and EARL T. APFEL. First published in "Iowa Geol. Survey," Vol. XXXIV (1928). Pp. 304; figs. 63; pls. 3; Pl. I in pocket. Part II: "The Illinoian and Post-Illinoian Pleistocene Geology of Iowa," by GEORGE F. KAY and JACK B. GRAHAM. First published in "Iowa Geol. Survey," Vol. XXXVIII (1943). Pp. 262; figs. 89. Part III: "The Bibliography of the Pleistocene Geology of Iowa," by GEORGE F. KAY. [1943]. Iowa City, Iowa: Iowa Geological Survey, [1943]. Pp. 55.

This volume is labeled *Special Report* because it combines parts of two volumes of the Iowa Geological Survey, to which is added "The Bibliography of the Pleistocene Geology of Iowa." As a special volume it constitutes a monographic treatment of the Pleistocene geology of Iowa.

In the Introduction to Part II, "The Illinoian and Post-Illinoian Pleistocene Geology of Iowa," the authors call attention to the fact that certain advances had been made since the publication of the report on "The Pre-Illinoian Pleistocene Geology of Iowa" in 1928—especially with respect to the classification of the Pleistocene—and they set forth the classifications shown in Tables 1 and 2 as those which are now recognized by the Iowa Geological Survey, one for the (upper) Mississippi Valley and the other for the state of Iowa.

If the reader bears in mind that fifteen years intervened between the publication of Parts I and II and that in the meantime revisions of the classification were made by Kay and Leighton, the principal discrepancies between Parts I and II will be clarified. The Iowan glacial stage had become a part of the Wisconsin, and the Peorian interglacial stage had become an *intra*glacial substage in the Iowa section. Furthermore, Kay had chosen to recognize the Nebraskan (glacial) and Aftonian (interglacial) as constituting an epoch to which he applied the name "Grandian," the Kansan (glacial) and Yarmouth (interglacial) another epoch which he called "Ottumwan," the Illinoian (glacial) and Sangamon (interglacial) a third epoch named "Centralian," and the Wisconsin (glacial) and Recent a fourth epoch labeled "Eldoran." Accordingly, the chapter headings of Part II are different in form from the corresponding chapter headings of Part I.

The reviewer has followed closely for thirty years the work which Kay has pursued so vigorously on the Pleistocene geology of Iowa. In 1911, upon the death of his predecessor, the late Professor Samuel Calvin, who had been state geologist almost continuously since the organization of the present Survey, Kay began his Pleistocene studies in Iowa while the writer was a student at the University of Iowa. Later, when the writer was assisting Dr. William C. Alden in a review of the evidences of the Iowan drift in northeastern Iowa, Kay invited them to participate in a field conference in southern Iowa to observe his findings with reference to the nature and origin of the "gumbo" which lay at the surface of the Kansan drift. Still later, after the writer of this review had become state geologist of Illinois, frequent interstate conferences were held over a period of twenty years.

Not a single field season was missed by Kay. He carried his investigations in a systematic manner from area to area and from problem to problem until he had covered the entire state and the whole Pleistocene section, including the bedrock topography upon which the glacial deposits rest. During each college year he dedicated himself to the other exacting responsibilities of professor, head of the department of geology, dean of the College of Liberal Arts, director of the State Geological Survey, and scientific scholar and writer. Nevertheless the monograph now under our review portrays a patient documentation of the features and phenomena observed, a critical and fair-minded review of the work and views of his predecessors and contemporaries, an astute analysis of the problems, a clear presentation of his own interpretations and concepts, and a frank recognition of baffling problems. At some points in the report the reviewer finds himself desiring to observe the field evidence in the light of his own

experience or from his own point of view, but nowhere does he question the devotion of the investigator to his problem.

That Iowa contains a rich record of the Pleistocene period and that Kay expended great energy in unraveling it will impress the readers of this outstanding volume. By his work the

TABLE 1

CLASSIFICATION OF PLEISTOCENE GEOLOGY IN THE MISSISSIPPI VALLEY

Period (System)	Epoch (Series)	Age (Stage)	Subage (Substage)
Pleistocene or Glacial	Eldoran	Recent Wisconsin	Mankato Cary Tazewell Iowan
	Centralian	Sangamon Illinoian	
	Ottumwan	Yarmouth Kansan	
	Grandian	Aftonian Nebraskan	

fundamental stratigraphy of the Pleistocene has been made secure. Hence this complete report has a special place in geological literature and is warmly received.

The discussion of each of the glacial drifts and interglacial phenomena is meticulous from both field and laboratory standpoints. In the case of each glacial drift, the criteria that may be used in its discrimination are set forth, followed by a discussion of its distribution, origin, changes it underwent subsequently, and typical sections in various areas and in different stratigraphic relationships. Then there follows a thoroughgoing description of the subdivisions of the whole profile of weathering, including the unweathered parent-material.

An unusual feature of the report is the presentation of laboratory results of sedimentation studies from the physical and chemical standpoints. The tills, silts, sands, and gravels and their weathered zones were sampled and subjected to careful laboratory testing and analysis by the junior authors and advanced students. This phase of study included their clastic texture, size-grade distribution, lithology, roundness of pebbles, and chemical composition. The resultant data were then assembled, compared, and contrasted to provide valuable information on the original constitution of the glacial-laid, water-laid, and wind-laid materials, on the changes they underwent by weathering in different zones of the weathering profile, and on the value of various criteria.

Considerable attention was given by Cornelia Cameron to the molluscan faunas of the Loveland loess and the Peorian loess, and by George H. Lane to the climatic and floral interpretation of peat zones, based on pollen grains. Further work of the foregoing character will contribute valuably to our knowledge and understanding of the sediments of the Pleistocene and to the conditions which prevailed. The mammalian and other vertebrate record contained in this report was derived mainly from the literature, but Kay makes a valuable contribution in pointing out that the stratigraphic position of a considerable number of the sources is open to serious question and that this field of knowledge should be thoroughly reviewed and modernized to provide a proper interpretation of the prevailing climatic and ecologic conditions.

Part I is devoted to separate chapters on the bedrock surface, topography and drainage, a review of investigations and classification of the Pleistocene deposits, the Nebraskan glacial stage, the Aftonian interglacial stage, the Kansan glacial stage, and the Yarmouth interglacial stage, together with an admirable sum-

TABLE 2

CLASSIFICATION OF THE PLEISTOCENE GEOLOGY OF IOWA

Period (System)	Epoch (Series)	Age (Stage)	Subage (Substage)
Pleistocene or Glacial	Eldoran	Recent Wisconsin	Mankato Peorian Iowan
	Centralian	Sangamon Illinoian	
	Ottumwan	Yarmouth Kansan	
	Grandian	Aftonian Nebraskan	

mary under the heading of "Concluding Statements."

The viewpoint is expressed that the preglacial topography of Iowa is by no means the present topography of the bedrock surface, either beneath the drift or in the so-called "Driftless Area" of northeastern Iowa. Kay

and Apfel point out that the present relief of northeastern Iowa is 200–400 feet greater than the relief of the bedrock surface under the glacial drift. This greater relief is thought to have been developed following the derangement of drainage by the Nebraskan glacier, which diverted the master-stream to the course of the present Mississippi River. In this way deeper dissection was initiated and accomplished in northeastern Iowa during the Aftonian interval. This concept of deep Aftonian erosion in northeastern Iowa is that of Trowbridge, who found that the Nebraskan drift there consists of scattered patches whose lower limit is about that of the Lancaster erosion surface. The authors also point out that the plotting of the average of the highest elevations of the bedrock in all parts of the state reveals that a topographic low—a reduced drainage basin—extended across Iowa from north to south, entering the state a little east of the middle of the north line and leaving it in the southeast corner.

From this knowledge they derive the concept that, at the close of the Tertiary, Iowa possessed a subdued topography near or at the Lancaster erosion level and that the incised valleys of northeastern Iowa and the steep-walled, drift-buried valleys in the Kansan drift area and some elsewhere were carved chiefly during the Aftonian interglacial stage. These concepts will await testing when sufficient well data and other information are in hand.

With respect to the present topography and drainage of Iowa, Kay and Apfel furnish a clear general picture of the variety of topographic conditions prevailing in the state. Among the points of outstanding interest are the Pleistocene age of the deep dissection of the so-called "Driftless Area" in northeastern Iowa; the mature stage of erosion of most of southern Iowa, involving successive grade levels, and more advanced erosion from east to west toward the Missouri River; the disappearance of tabular divides on the Kansan drift of northwestern Iowa from south to north; the drift-mantled erosional topography of the Iowan drift area in eastern Iowa and its border of loess-mantled erosional topography; the unique loess depositional topography of western Iowa bordering the Missouri flood plain, changing eastward into a belt of loess-mantled erosional topography; and the youthful glacial features of the drift-depositional topography of the Des Moines lobe.

Chapter iii of Part I gives a thorough and instructive review of the history of investigations of the Pleistocene deposits of Iowa, interwoven with the development of the classifications. C. A. White, in 1870, first recognized the glacial origin of the surficial materials in Iowa, but McGee made the first important contributions to the unraveling of the glacial history of the deposits in 1878. Chamberlin had already differentiated in a masterly manner two drifts in Wisconsin and established the record of an interglacial epoch, as had Winchell in Minnesota.

Kay and Apfel review the comprehensive work of McGee in a spirit of understanding of the state of knowledge of that time and without criticism of his failure to recognize three tills instead of two. They emphasize, instead, his full treatment, for that time, of the Upper Till and the Lower Till of northeastern Iowa and its associated loess, subsequently named "East Iowan" and then "Iowan," and "Kansan," respectively; his recognition of the youth of the Upper Till compared with the gumbo-surfaced drift (present Kansan) of southern Iowa and its greater age than the Des Moines lobe (present Wisconsin). Had McGee recognized the presence of three tills, both Chamberlin and McGee might have viewed the Afton Junction section in a different light when they visited it in 1893, made proper correlations with the tills of northeastern Iowa, and saved later controversy regarding nomenclature.

These miscorrelations were discovered by Bain and recognized by Chamberlin in a field conference in 1896, and because the name "Kansan" had in the meantime been applied to the uppermost drift in areas bordering and lying outside the Iowan drift, Chamberlin acceded somewhat hesitantly to the designation of the upper till at the Afton Junction section as Kansan instead of the lower till. Then in his editorial that autumn in the *Journal of Geology* he applied the name "Albertan" (after Dawson) to the lower till. It seems difficult to understand, from this distance, how the upper till at Afton Junction, with loess overlying it unconformably, could have been correlated with McGee's Upper Till (Iowan), inasmuch as McGee had regarded the Upper Till and the loess as of the same age.

The organization of the Iowa Geological Survey in 1892 came at an opportune time for Calvin, Norton, Bain, Udden, Shimek, Tilton, Beyer, and others to pursue vigorously Pleistocene and other studies during the rest of the

1890's and the early 1900's, when knowledge of the Pleistocene was rapidly developing. By 1896, Leverett, of the United States Geological Survey, had discovered the Illinoian stage of glaciation in Illinois and southeastern Iowa, and Shimek had set forth his eolian theory of origin for the loess deposits. By 1898 the names of the principal glacial and interglacial stages had been virtually agreed upon for the Pleistocene of the upper Mississippi Valley by the leading students, with the exception of the Nebraskan, which was suggested by Shimek in 1909. Thus the youthful Iowa Geological Survey shared in this very constructive scientific period.

The controversy that ensued later with regard to the Iowan drift is thoroughly described and analyzed by the authors without prejudice. As a result of this controversy and the additional critical studies made, the Iowan drift and its relationships came to be more fully known and appreciated for its special features.

The first contributions of Kay in regard to the nature and origin of the gumbo on the tabular divides of the Kansan drift of southern Iowa began to appear in 1916; and, as the reader follows the history of his and Apfel's investigations of the gumbotil of the Nebraskan, Kansan, and Illinoian drifts, he cannot help realizing that this later period of refined and detailed studies marks another constructive period for Pleistocene geology. The Aftonian, Yarmouth, and Sangamon interglacial horizons in Iowa became as firmly established as any other part of our whole geologic column, and the use of the Kansan and Nebraskan gumbotil horizons made possible the areal mapping of the Nebraskan drift over a large part of southwestern Iowa and the general contouring of the original surfaces of the Nebraskan and Kansan drift over a large part of the state. Perhaps less notable but yet important was Kay's work on the Loveland formation and the Loveland loess, the latter of late Sangamon age. The position of the Loveland loess above the Illinoian drift and below the Iowan drift is one of the conclusive criteria for fixing the age difference of these two drift sheets and for fixing the upper limit of the Sangamon interglacial horizon.

It will pay the young student of the Pleistocene to read thoroughly this chapter on the history of investigations and of the classifications of the Pleistocene of Iowa, because, in addition to setting forth the subjects mentioned above, it covers many other important features, including the work on the Iowan drift of northwestern Iowa by Carman, the fossil-bearing gravels of southwestern Iowa by Calvin and Shimek, the Pleistocene mammals of Iowa by Hay, the history of the Des Moines River Valley by Lees, the history of Lake Calvin by Schoewe, and several other special studies.

Illuminated by the authors' "History of the Investigations and Classifications of the Pleistocene Deposits of Iowa," the chapters that follow constitute a thoroughgoing treatise of present knowledge of the stratigraphy and history of the Nebraskan glacial stage, the Aftonian interglacial stage, the Kansan glacial stage, and the Yarmouth interglacial stage.

These chapters are methodically organized. The Nebraskan and Kansan drifts are described and discussed with respect to their stratigraphic relations, distribution, origin, changes, typical sections, drift phases (profiles of weathering), and thickness.

The chapter on the Aftonian discusses the record of that interval, sections representative of it, the Nebraskan gumbotil, erosion during the interval, Aftonian loess, and the life of that epoch.

The chapter on the Yarmouth includes a statement of the record; a discussion of the Buchanan interval (post-Kansan, pre-Iowan); descriptions of the weathered products of the Kansan drift beneath the Illinoian drift and outside the Illinoian area; the buried soils and vegetal materials; erosion; silts and sands; the Loveland formation; and the record of life in the Yarmouth.

The authors point out that the Nebraskan drift in many places "can be identified with certainty only if one can establish definitely its relations to certain interglacial materials the age of which is known." The Nebraskan ice, from the Keewatin center, covered all of Iowa; and when it melted, it left drift of sufficient thickness and so distributed over the bedrock surface as to produce a plain with minor surface irregularities. The probable average thickness of the drift for the state is estimated at more than 100 feet. It is difficult to distinguish it from overlying Kansan lithologically except in northwestern Iowa, where it contains a large amount of Cretaceous shale material. The Kansan ice, also from the Keewatin center, covered the state, with the exception of the so-called "Driftless Area," and deposited a massive sheet of drift over the gumbo-surfaced and eroded Nebraskan, leaving a strikingly level or nearly level ground-moraine plain.

It is emphasized that the gumbotils are the most satisfactory horizon markers in the state. It is believed that, when each was formed, the ground-moraine plains were low lying and that dissection of these plains was subsequently initiated by uplift. The reviewer, without denying that there may have been uplift, is of the opinion, after making an extensive study of the Illinoian drift plain in Illinois, that the assumption of uplift is not necessary because the essential topographic condition for the formation of gumbotil is a degree of flatness essential to poor surface and subsurface drainage. Gumbotil is found on the Illinoian drift on plains lying at different levels, but plains sufficiently extensive to have poor drainage.

The authors note from their studies that the so-called "Aftonian gravels" are not of Aftonian age but of Nebraskan, some of which were at the surface and became weathered before the Kansan ice invasion. They suggest that they be called "weathered Nebraskan gravels" and that the name "Aftonian gravels" be dropped. Likewise the sands and gravels of western Iowa are contemporaneous in age with the tills with which they are associated.

The numerous outcrops of gumbotil under the Kansan drift in southwestern Iowa indicate that extensive valleys there of Aftonian age were few. Elsewhere erosion was great, which is especially clear in northeastern Iowa.

Thin loess, referred to the Aftonian, is present in Lyon and Union counties, and thin "loess-like clay" in Shelby County.

As noted hitherto, Part II was prepared several years subsequent to Part I, and Kay's junior associate was Jack B. Graham. This part bears the mark of Kay's new classification in which he combines the Nebraskan and Aftonian to make the Grandian epoch, the Kansan and Yarmouth the Ottumwan epoch, the Illinoian and Sangamon the Centralian epoch, and the Wisconsin and Recent the Eldoran epoch, all of them comprising the Pleistocene period. R. T. Chamberlin has taken exception to ranking the Pleistocene as a system or period for reasons which he set forth with impressive clarity in a recent number of the *Journal of Geology*.

As for the pairing of a glacial stage and the immediately following interglacial stage to form an epoch, the reviewer believes that there is a degree of justification for this in that the altered and secondary materials of the weathered zone of a drift sheet, where overlain by the next younger drift, can then be assigned to the epoch to which they belong, being partly of the glacial age and partly of the interglacial age.

In accordance with this new grouping the chapter headings of Part II run as follows: chapter i, "The Centralian Epoch (Series): The Illinoian Glacial Age (Stage)"; chapter ii, "The Centralian Epoch (Series): The Sangamon Interglacial Age (Stage)"; chapter iii, "The Eldoran Epoch (Series): The Wisconsin Glacial Age (Stage)"; chapter iv, "The Eldoran Epoch (Series): The Recent Interglacial Age (Stage)." The treatment of the glacial drifts and interglacial deposits is, on the whole, similar to that followed in Part I.

Part II also introduces the feature of profiles of weathering, which subject had its development in this country mostly subsequent to the publication of Part I. In this period Leighton and MacClintock amplified and extended the concept of the gumbotil profile to include the "silttil profile," developed under conditions of good drainage in rolling topography, and the mesotil profile, developed under conditions intermediate between good drainage and poor drainage. They also introduced the concept of second-cycle profiles resulting from headward erosion changing poor drainage conditions to good drainage and permitting the colloids of gumbotil to be carried out to form the product silttil (second-cycle).

On page 13 of Part II, Kay and Graham present two new classifications, one for the Mississippi Valley and the other for Iowa. The latter differs from the former in that two of the Wisconsin substages—Tazewell and Cary—are not represented in Iowa by glacial deposits but by Peorian loess. Hence the name "Peorian" is substituted in the Iowa section for Tazewell and Cary. In this connection it is well to draw attention to the fact that the Peorian loess exposed beneath the eastern edge of the Mankato drift of Iowa has a shallow zone of leaching, thus recording that the Peorian substage here is represented both by deposition of loess and by weathering. It is also well to remember that the loess along the immediate eastern and southern borders of the Iowan drift is mostly late Iowan in age, having been blown from the exposed fringe of bare-surfaced Iowan drift while the Iowan ice was still melting away. Along the Mississippi and Missouri rivers there may also be loess of Tazewell, Cary, and Mankato ages as well as Iowan, which had their source from valley-train silts of those times.

The Pleistocene student will find special in-

terest in the discussion of the Loveland formation and its significance, in the well-rounded data concerning the Iowan drift, and in the full treatment of the Peorian loess, including mechanical, mineral, and chemical analyses and the paleontology and ecology of the molluskan fauna and flora. Greater emphasis should have been placed, the reviewer believes, upon the original content and nature of the Peorian loess deposits. The authors cite certain sections of Peorian loess exposed beneath Mankato drift at the southern edge of the Des Moines lobe as "unusual" because there is much woody material within the body of the loess. Such sections are unusual in the sense that they are comparatively few, but they have special significance. The reviewer has found that woody material is common in sections of Peorian loess in Illinois where the loess is buried and preserved from weathering by the overlying Wisconsin drift. The prevailing conception of loess comes from the widespread exposures of loess that has been oxidized and, through the oxidation, has lost its original woody content. A proper conception of the original nature of the Peorian loess of the upper Mississippi Valley is to be derived from a study of those sections which have been preserved from weathering by an overlying cover of younger Wisconsin drift.

Part III, "The Bibliography of the Pleistocene Geology of Iowa," was prepared by Professor Kay with the assistance of several graduate students. This excellent addition to this monograph is a valuable source of reference not only for students of the Pleistocene of Iowa but for those working in bordering states or in special phases of the Glacial period of North America.

It is fortunate, indeed, that the junior authors of Parts I and II were privileged to be associated with the senior author in the preparation of this special report on *The Pleistocene Geology of Iowa*. Such a relationship as this which Professor Kay made possible carried with it his acknowledged credit of the important contributions that they made. This adds more than the reviewer can add to all those attributes which lie behind the personality of the senior author, who fortunately was able to see this work in final proof form before he was taken from his international family of scientific friends.

MORRIS M. LEIGHTON